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EARLY DIFFERENTIATION OF THE MOON: EXPERIMENTAL AND MODELING STUDIES

and

EXPERIMENTAL AND MODELING STUDIES OF MASSIF ANORTHOSITES

3/1/89 - 8/31/94

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Summary

NASA grant NAG9-329 was in effect from 3/1/89 to 8/31/94, the last 18 months being a no-cost extension. During the period of the grant John Longhi was P.I. and held the positions of Associate Research Scientist, Research Scientist, and Senior Research Scientist at the Lamont-Doherty Earth Observatory of Columbia University. While the grant was in effect, the P.I., co-workers, and students gave 22 talks and poster sessions at professional meetings, published 12 articles in refereed journals (one more is in press, and another is in review), and edited 2 workshop reports relevant to this project. Copies of all the publications are appended to this report.

The major accomplishments during the grant period have derived from three quarters: 1) the application of quantitative models of fractional crystallization and partial melting (Longhi, 1991a) to various problems in planetary science, such as the petrogenesis of picritic glases and mare basalts (Longhi, 1992a and b) and the implications of the SNC meteorites for martian evolution (Longhi, 1991b; Longhi et al., 1992); 2) an experimental study of silicate liquid immiscibility relevant to early lunar differentiation and the petrogenesis of evolved highlands rocks (Longhi, 1990); and 3) experimental studies of massif anorthosites and related rocks that provide terrestrial analogs for the proposed orgin of lunar anorthosites by multi-stage processes (Fram and Longhi, 1991; Longhi et al. 1993; Vander Auwera and Longhi, 1994). The low-pressure aspects of the quantitative models were developed by the P.I. in the 1980s with NASA support and culminated with a paper comparing the crystallization of terrestrial and lunar lavas (Longhi, 1991a). The basis for the high-pressure modifications to tbe quantitative models is a data set gleaned from high-pressure melting experiments done at Lamont (Longhi, 1993; Longhi, 1994; Vander Auwera and Longhi, 1994) as shown in Fig. 1 and is supplemented by published data from other labs that constrain the baric and compositional dependences of various liquidus phase boundaries, such as olivine/orthopyroxene, relevant to the melting of the mantles of the terrestrial planets. With these models it is possible to predict not only the thermal and compositional evolution of magmatic liquids ranging in composition from lumar mare basalt to terrestrial calc-alkaline basalts, but also the small increments of fractional melting that are produced when mantle rises adiabatically. Copies of the crystallization/melting programs have been given to several colleagues in planetary science. Additionally, a series of computer graphics programs, based on the algorithms in the crystallization programs, have been developed that display liquidus diagrams appropriate to input compositions (Longhi, 1991a).

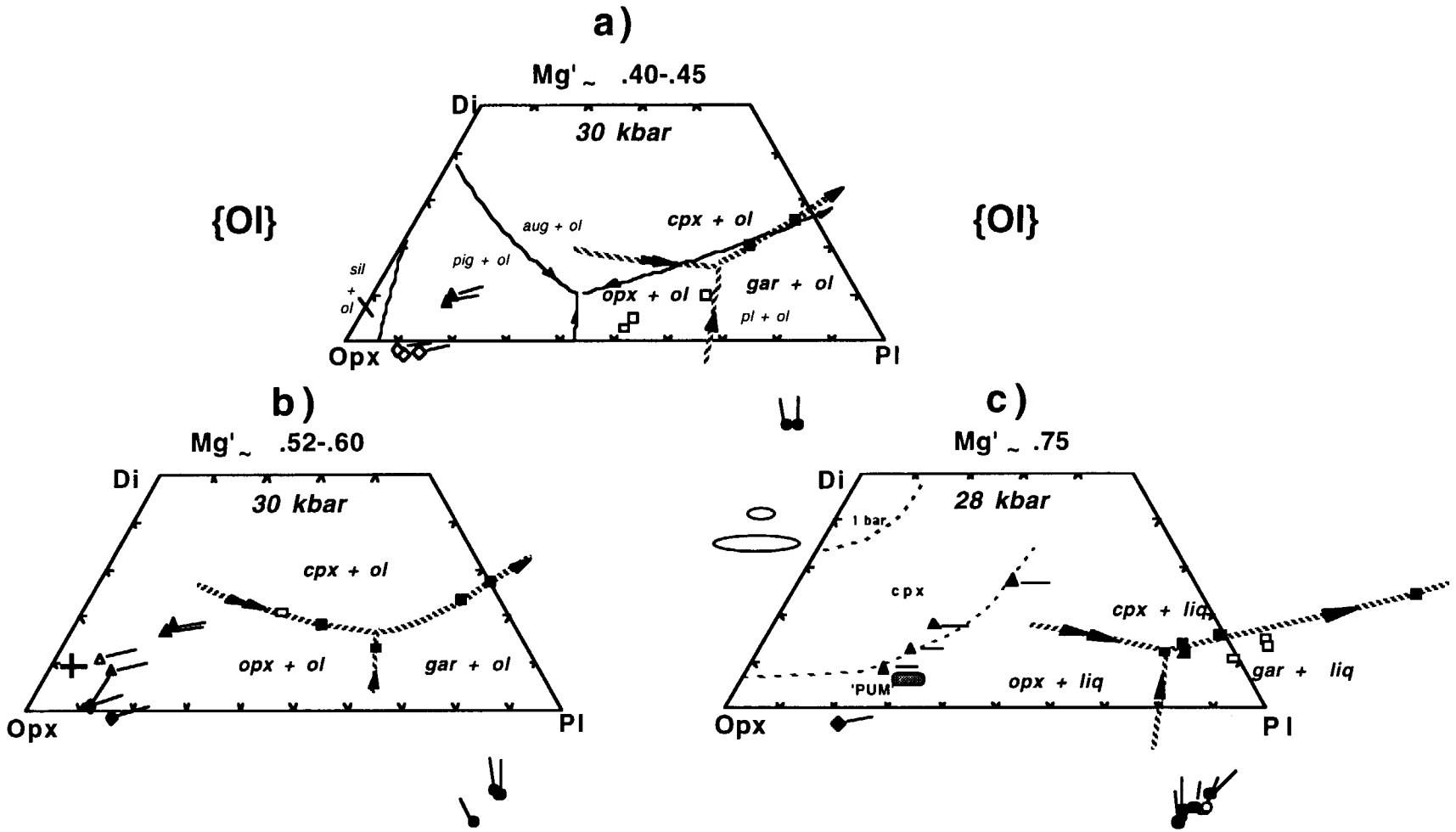

Figure 1. Projections from Olivine component {Ol} onto a portion of the Wollastonite(Wo) - Orthopyroxene(Opx) - Plagioclase(Pl) plane. Olivine coexists with all phases: a) $fo_{67\text{-}72}$, b) $fo_{75\text{-}83}$, c) $fo_{87\text{-}91}$. Light dashed curves show limits of clinopyroxene solid solution. light solid line in c) is tieline joining coexisting opx and cpx. Heavy dashed curves are inferred high-pressure liquidus boundaries. Squares = liquid compositions; circles = garnet; triangles = cpx; diamonds = opx. Filled symbols indicate coexistence of 3 solid phases plus liquid (e.g., ol+opx+gar+liq); open symbols indicate coexistence of 2 solid phases plus liquid. Arrows show direction of decreasing temperature; double arrows show reaction boundaries. Large ellipse in c) shows range in composition of liquid analyses (~30 micron raster of quench crystals) in a single run; smaller ellipse shows compositional range of pyroxene analyses (spot). Cross in b) is estimated green glass magma source composition from Longhi (1992a); "PUM" field shows estimates of terrestrial depleted and undepleted upper mantle compositions (Kinzler and Grove, 1992).

Picritic Glasses/Mare Basalts

Work in this area has centered primarily on developing an alternative model for the petrogenesis of mare picritic glasses (Delano, 1986). As their description implies, these glasses have high contents of normative olivine. These high olivine contents in turn produce high-pressure liquidus multiple saturation of olivine and pyroxene in the range of 17 to 25 kbar. The standard interpretations are either that this pressure range is the minimum pressure range of melt segregation from a multi-phase residuum or that there were sufficiently high degrees of partial melting to exhaust pyroxene from the source at a lower pressure. On the Moon 25 kbar implies depths in excess of 500 km and the prospect of transporting a magma that distance without fractionating it seems dubious. One calculation suggests magma traveling 10 m/sec in a through-going crack 40 m wide is necessary to avoid crystallization (Spera, 1992). Alternatively, a cumulate source with 2.0 wt % Al_2O_3 and 80 % olivine would require more than 30 % partial melting; sources with more Al_2O_3 or less olivine would require even higher degrees of melting. Polybaric melting models provide one way out of this dilemma. As described by (Klein and Langmuir, 1987), melting and segregation takes place over a range of depths with the melt accumulating into a pool that eventually finds a crack and makes it to the surface. Thus melting experiments give only an average pressure of segregation: melting begins a higher pressures and continues to lower pressures.

Fig. 2 from Longhi (1992a) illustrates traces of the pooled melt composition for polybaric fractional melting of a model mare source. The origins of the curved lines in Fig. 2a are the compositions of the intial melts at the initial pressure of melting. The tips of the arrows are final pooled melt compositions when the melting stops at 5 or 6 kbar. The source composition and solidus changes in response to melt segregation in these calculations, so some melt must be retained in the source to keep the source from exhausting its basaltic components and to keep the temperature required to melt the depleting source from rising above an adiabatic gradient. The letters adjacent to the curved paths are keyed to the calculation parameters in the inset table: EF is the fraction of melt generation in each step that is withdrawn in each step. The melting paths that begin at 40 kbar are quite successful compositionally and imply great depths (> 1000 km) of differentiation and, possibly, initial melting in the Moon. Revised calculations, employing modified equations for the pressure dependence of liquidus boundaries and partion coefficients (Longhi, 1993 and 1994). The major challenge to this model is the physical requirement of maintaining a distinct pool of melt that can be fed by the partially melting source region as it ascends.

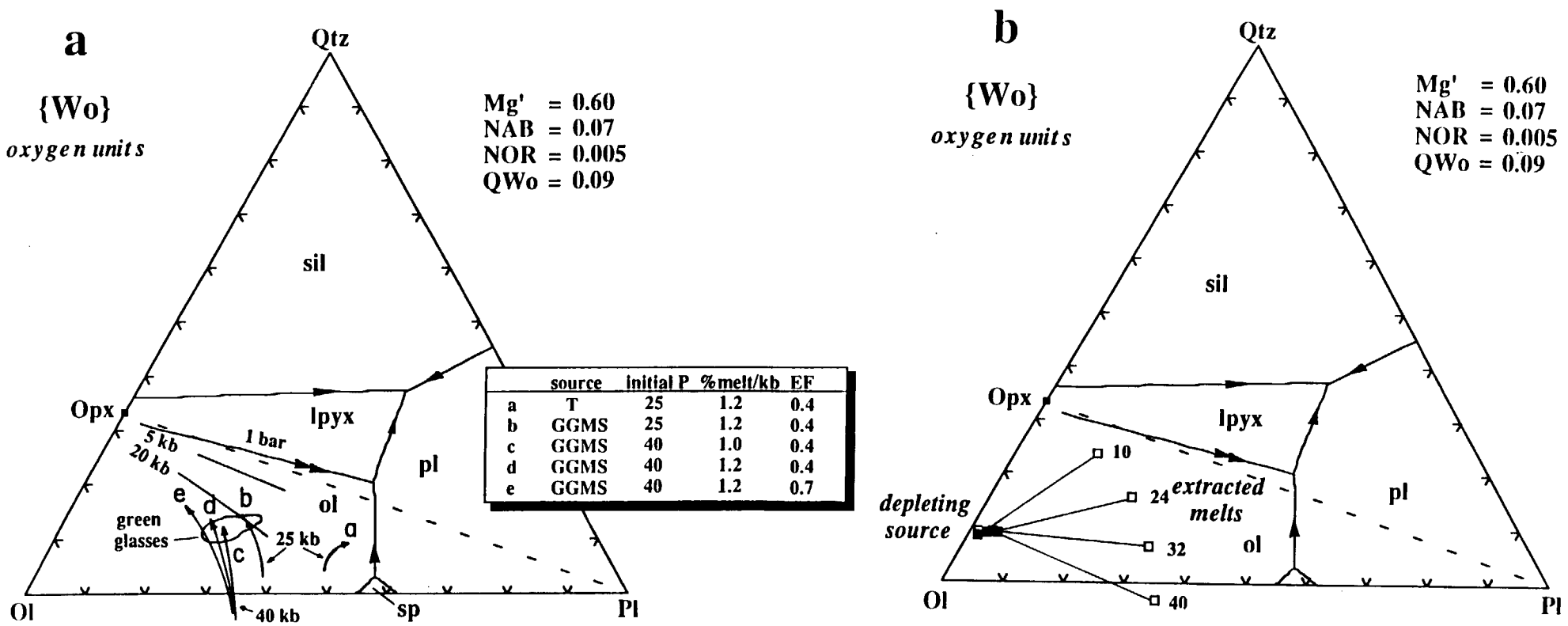

Figure 2. Phase equilibria and compositional relationships of green glass and calculated melts shown in terms of the projection from the Wollastonite component {Wo} onto the Olivine (Ol) - Plagioclase (Pl) - Quartz (Qtz) plane. (a) Comparison of green glass compositions (Delano, 1986) and calculated pooled melt compositions derived by polybaric fractional fusion. Low-pressure (1 bar) phase boundaries are calculated from the liquidus boundary equations in Longhi (1991a) and are appropriate for green glass compositional parameters shown in upper right where Mg' = MgO/(FeO+MgO) - molar, NAB and NOR are the albite and orthoclase fractions of normative feldspar, and QWo is the quaternary coordinate of the Ol-Pl-Wo-Qtz model system in mole units. Lower case letters are keyed to inset table where T refers to the average mantle composition of (Taylor, 1982), GGMS is green glass model source composition (see Table 2 in Longhi (1992a), EF is melt extraction efficiency (percentage of melt extracted from the source in each cycle). Phase abbreviations: ol = olivine, sil = silica phase, lpyx = low-Ca pyroxene (orthopyroxene or pigeonite), pl = plagioclase, sp = spinel. b) Series of instantaneous partial melts joined to source composition from which they were generated. The instantaneous melts have distinct compositional parameters (Mg', NAB, etc.) from one another and, in general, from pooled melts or green glasses, so that the pressure dependence of the liquidus boundaries, which are compositionally dependent, may not be simply gauged from the figure.

SNC Meteorites and Mars

Work in this area has involved investigation of the patterns of incompatible element and isotopic fractionations in the shergottite-nahklite-chassignite (SNC) parent magmas (Longhi, 1991b; Longhi, 1992c). An important consequence of this effort was the discovery that several incompatible high-field strength elements (HFSE) are fractionated relative to the rare earth elements (REE) as shown in Fig. 3; and there is also a Sr "signal" at least in the Antarctic shergottites. The order of elements chosen in Fig. 3 is a modification of the standard "spider diagram" that removes volatile elements to facillitate interplanetary comparisons of melting processes and emphacizes the REE. A search through the literature shows that Sr anomalies are also common in lunar, ocean island (OIB), and convergent plate margin basalts. In the lunar case relative depletions in Sr correlate well with relative depletions of Eu -- the well-known signature of plagioclase fractionation. However, this not the case for the SNC magmas or for primitive terrestrial basalts that show weak or absent Eu anomalies and marked enrichments of Sr. Also, in many cases these positive Sr anomalies correlate well with apparent positive anomalies in Ba. Because Sr and Ba are non-volatile incompatible elements readily transported by H-C-O fluids (e.g., Gill, 1981) and because these two elements may be enriched even when the light REE are depleted relative to the heavy REE, it seems possible that fluids have played a role (metasomatic?) in the petrogenesis of SNC parent magmas as well as in terrestrial island arc and ocean island basalts.

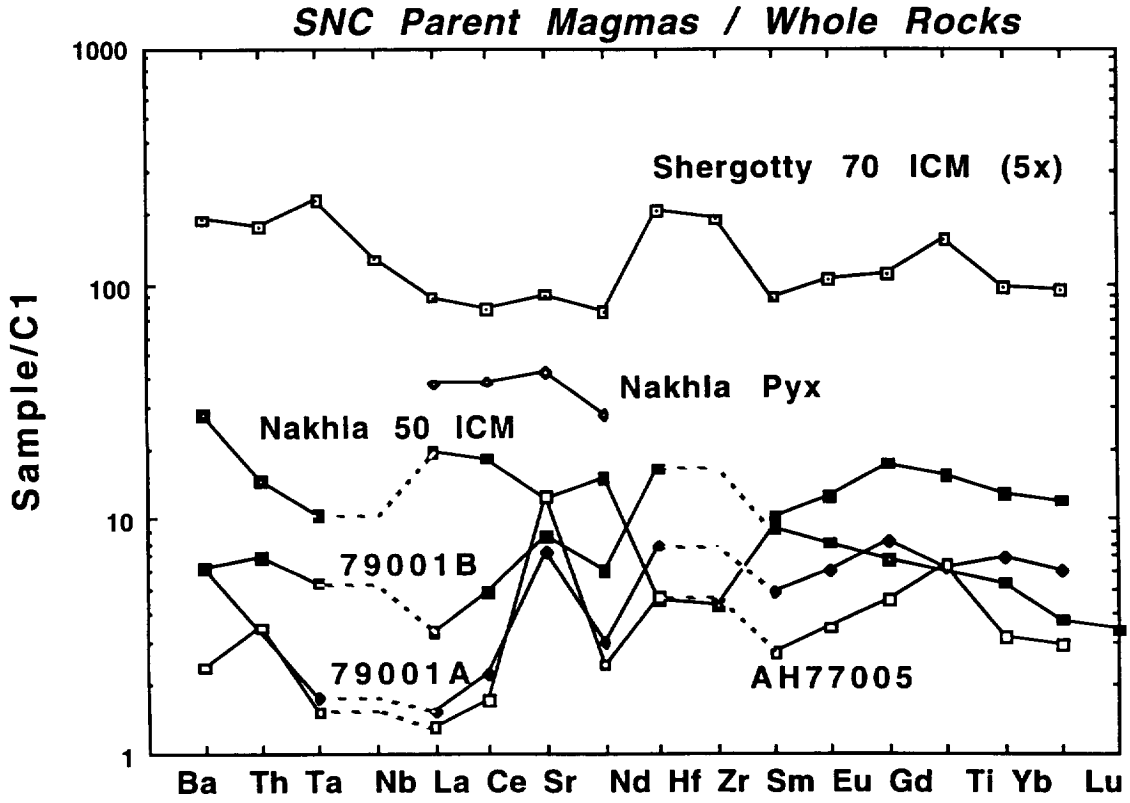

Fig. 3. Chondrite-normalized non-volatile incompatible element patterns for SNC parent magmas. Elements are arranged in order of increasing incompatibility left to right. ICM refers to assumed amount of intercumulus melt in whole rock. Pyx refers to magma composition calculated from pyroxene separate.

Silicate Liquid Immiscibility

Silicate liquid immiscibility (SLI) remains a frequently invoked mechanism for producing chemical fractionations observed in some lunar highlands rocks. There is also petrographic evidence of SLI in highland breccias. Experimental work in this lab has shown that SLI is stable in plagioclase-saturated lunar liquids under isothermal conditions and has led to the development of a crude empirical algorithm predicting the size of the two-liquid field in evolved liquids (Longhi, 1990). Results of calculations with this algorithm show that SLI is likely to develop during the late stages of post-ferroan-anorthosite primordial differentiation (*ur*KREEP), during extreme fractionation of the magmas that produced the high-Mg pristine cumulates, and also in the late stages of differentiation of KREEP basaltic magmas. These calculations demonstrate that both SLI and KREEP-like enrichments of incompatible elements would have developed in post-ferroan-anorthosite residual liquids only after the liquids became hyperferroan. Thus granitic highland rocks, which typically have mafics with intermediate Mg' (Warren et al., 1983), cannot be related to post-ferroan-anorthosite (primordial?) differentiation.

On a more mundane level the isothermal crystallization experiments in which SLI developed were all saturated with a silica mineral in contrast to the controlled-cooling-rate experiments of the Brown group. A second round of isothermal experiments on synthetic compositions with KREEP levels of Zr and REE were initiated to examine the discrepancy observed between controlled cooling rate experiments on natural compositions in which a silica phase did not crystallize (Hess et al., 1975; Rutherford et al., 1974) and isothermal experiments on similar synthetic compositions in which a silica phase did crystallize (Longhi, 1990). Silica is present in all of the new experiments in which SLI is observed, thus supporting the contention that the absence of silica in the SLI-bearing controlled cooling rate experiments was the result of a nucleation failure and not an equilibrium effect.

Massif Anorthosites

The goal of this project has been to study the seemingly polybaric petrogenesis of massif (Proterozoic) anorthosites as possible analogs of lunar anorthosites. What the two groups of anorthosites have in common is intermediate Mg/Fe in mafic minerals indicating that their parent liquids were probably evolved from more primitive magmas, a relatively limited range in plagioclase composition suggesting suspension and accumulation, and separation from their mafic cumulates that is at least on the order of crustal thickness. To be sure there are many significant differences — plagioclase composition, tectonic setting, and areal extent — but what is more important than the obvious differences is the development of tangible models of large scale anorthosite petrogenesis against which the limited contextural information but extensive mineral and chemical data base of lunar anorthosites can be compared. The incentive for developing terrestrial analogs has increased as diversity within the ferroan anorthosite suite has come to the fore (Warren et al., 1991) and as detailed geochemical models have suggested that ferroan anorthosites may not have crystallized directly from a homogeneous magma ocean (James et al., 1988). This latter situation is ironic: although the concept of a magma ocean was originally developed to explain existence of lunar anorthosites, it is possible, now that there is independent indication of the existence of a magma ocean formed by a giant impact, that lunar anorthosites may not have formed directly from a magma ocean, but rather from its aftermath. In any event, whether ferroan anorthosites were passively accreted to the bottom of a crust floating atop a magma ocean or whether ferroan anorthosites formed in plumes of plagioclase mushes detached from stagnant bodies of melt too dense to erupt on the surface, it seems prudent to investigate terrestrial anorthosites in parallel with lunar anorthosites even though terrestrial anorthosites may not prove to be exact analogs.

To date we have worked on the melting relations of hree compositions: a powdered rock sample of anorthositic dike from Nain proposed by Wiebe (1990) to be an anorthosite parent magma, a synthetic glass with the composition of the averaged high-Al gabbros that intruded the Harp Lake anorthosite complex and which Emslie (1980) had proposed as likely parent magmas, and a powdered rock sample from the chill margin of the Bjerkreim-Sockndal intrusion in southwestern Norway. The purpose of studying these compositions was threefold: first, to determine the plausibility of the three compositions as quenched liquids; second, obtain crystal/melt partition that would allow us to constrain the compositions of the liquids from which the anorthosites crystallized; and third, to generate multi-saturated liquids in the range of 5 to 15 kbar, as a basis for constructing liquidus phase diagrams that would help constrain the origin of the liquids parental to the massif anorthosites.

Results from (Fram and Longhi, 1992) show that the anorthositic dike has such a high liquidus temperature (>1350° C) at crustal pressures that it is unlikely this dike was ever completely molten. The plagioclase composition data tell a similar story: the near-liquidus experimental plagioclase (An_{60-67}) at upper crustal pressures are distinctly more anorthitic than the most anorthitic cores of the zoned natural plagioclase reported by Wiebe (1990) or analyzed at Lamont. Examination of all of the anorthositic dike compositions published by Wiebe reveals that there is wide scattering of compositions that cannot be attributed to addition or removal of plagioclase (Fig. 4). In fact some of the compositions, noritic anorthosite and troctolitic anorthosite, appear to be those of adcumulates. The overall anorthositic compositions of the dikes suggests intrusion of plagioclase-rich suspensions or mushes; the wide scatter in projected compositions indicates migration of the interstitial liquid relative to a matrix of plagioclase and mafic minerals. Thus, in general, the dike compositions do not represent simple suspensions of plagioclase in liquid.

Fram and Longhi (1992) showed that the high-Al basaltic composition (HLCA) has plagioclase, orthopyroxene, and augite coexisitng on the liquidus at ~12 kbar. At this pressure there is a good match in Mg/Fe and Al_2O_3 concentration between the most magnesian experimental and natural aluminous pyroxene megacrysts, traces of which are found in nearly every anorthosite massif. Fig. 5a compares the Al_2O_3 and CaO concentrations in megacryst and matrix opx from Harp Lake with those produced in isobaric experiements: there is apparently a strong dependence of $Al_2O_3$with pressure, but not CaO. The Al_2O_3 concentrations of the megacryst and experimental opx match at 11.5 kbar — the same pressure at which HLCA is cotectic. Fig. 5b shows a reversal of the Al_2O_3 concentration in the experimental opx (Longhi et al. (1993). Thus the pressure-dependence of the Al_2O_3 concentration is well-calibrated. Additionally, solute-rejection calculations by Longhi et al. (1993) using Cr, a compatible element, coupled with Al, an incompatible element, indicate that the positive correlation between high-Cr_2O_3 and high-Al_2O_3 concentrations observed in the megacryts are unlikely to be the result of rapid crystal growth or a failure of plagioclase to nucleate (e.g., Morse, 1982), because these kinetic processes deplete Cr in orthopyroxene while enriching Al . Fram and Longhi (1992) also showed that at upper crustal pressures, the liquidus plagioclase (interpolated) of HLCA is close in composition to the most calcic plagioclase reported at Harp Lake; however, at lower crustal pressures the liquidus plagioclase is similar to the most common plagioclase compositions at Harp Lake. This similarity is consistent with most of the plagioclase in the intrusion having crystallized at the same pressure as the opx megacrysts (~12 kbar) and thus having intruded upwards in a mush or suspension carrying the megacrysts along.

Mass balance indicates that as much as 80% plagioclase needs to be added to HLCA to account for the composition of the exposed portion of Harp Lake. Somewhat less plagioclase would be necessary, if the unexposed, lower portions of the complex were less anorthositic as at Michikamau (Emslie, 1970). Nonetheless, high proportions of crystals to liquid would still be required to keep the plagioclase from back-reacting to more calcic compositions as it intruded upwards. Longhi et al. (1993) have used contiguity theory (German, 1984) to argue that a suspension of up to ~ 70% plagioclase may maintain liquid, albeit highly viscous, rheolology

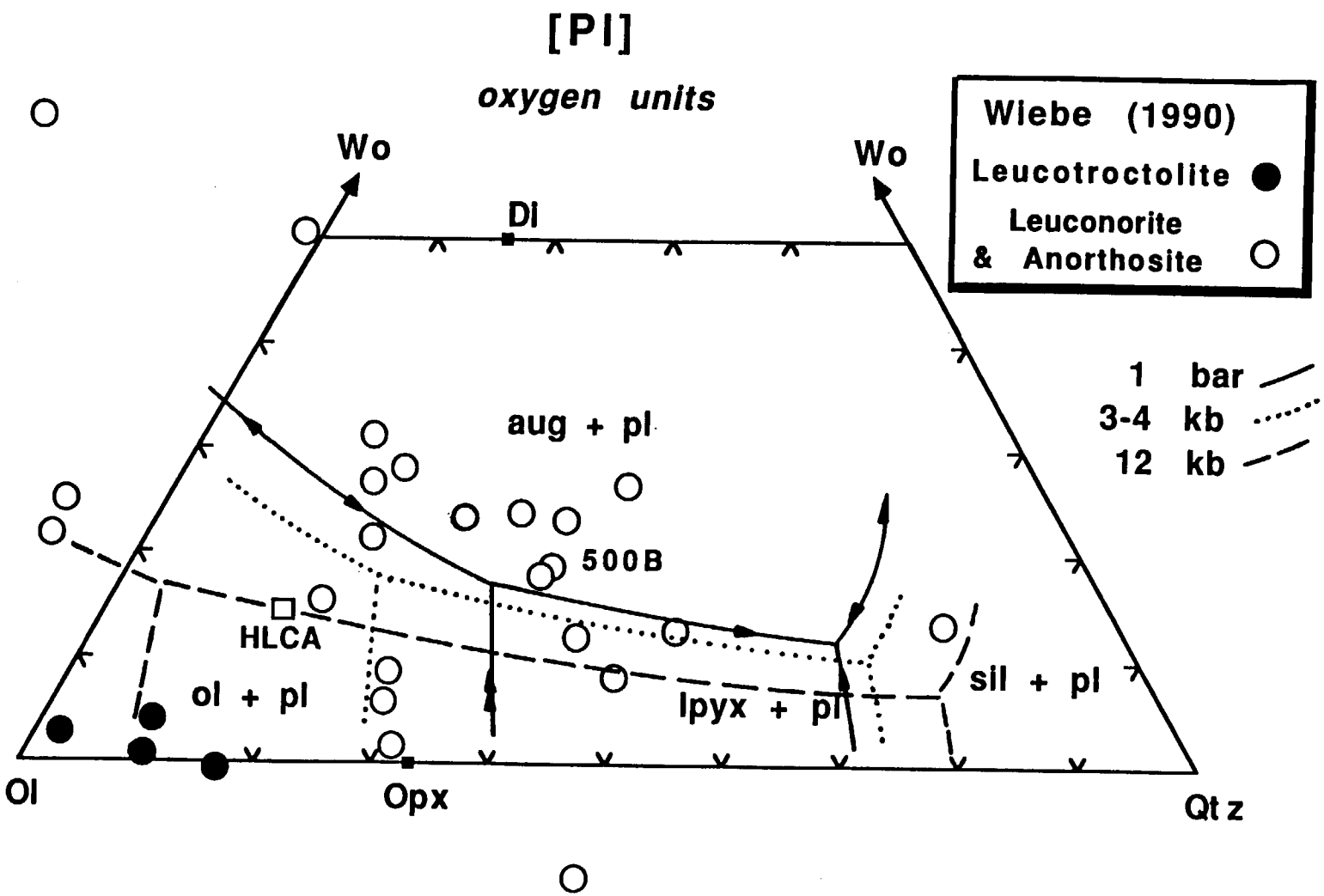

Fig. 4. Projection from plagioclase component onto the olivine-wollastonite-silica surface;1-bar and 12 kbar boundaries are based upon results of 500B and HLCA experiments; 3-kbar boundaries are interpolated. Liquidus field labels refer to 1-bar boundaries. Compositions of leucotroctolite dikes are shown as shaded circles, leuconorite and anorthosite dikes as open circles.

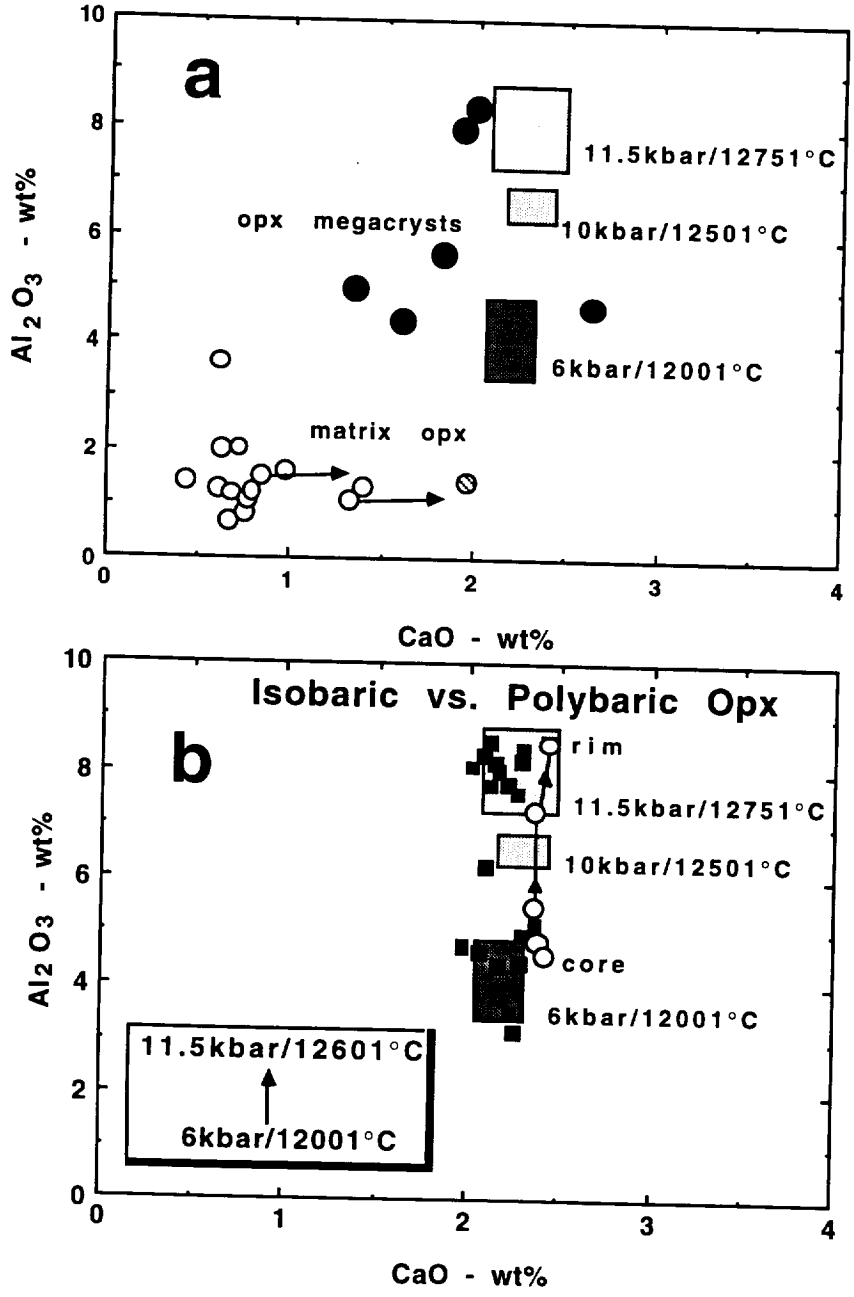

Fig. 5. Al_2O_3 and CaO concentrations (wt %) in orthopyroxenes. a) Open circles: spot analyses of exsolved matrix orthopyroxenes from the Harp Lake Complex; shaded circle: integrated composition of opx host with augite lamellae from Harp Lake; vectors indicate mixing paths between opx hosts and augite lamellae; solid circles: bulk and integrated opx megacryst compositions from Harp Lake and other massif anorthosite complexes; shaded boxes show one standard deviation about average opx compositions from experiments on HLCA. All natural compositions from Emslie (1980); experimental data from Fram and Longhi (1992). b) shaded boxes as in Fig. 3. Filled squares: random spot analyses of orthopyroxenes produced in polybaric experiments; open circles = core to rim traverses of single crystals with arrows indicating direction of rim.

(no deformation of the suspended crystals); whereas higher proportions plagioclase would be subject to solid state flow laws involving deformation of the crystalline matrix of plagioclase and thus incrreasing the viscosity very abruptly by several orders of magnitude. Taken together, the results of these investigations are consistent with a polybaric origin for massif anorthosites by upward intrusion of plagioclase-rich suspensions or mushes generated by floatation in deep-seated, magma chambers. Such scenarios were discussed for lunar ferroan anorthosites by Longhi and Ashwal (1984). However, a second stage of plagioclase enrichment would be needed to produce true anorthosites, both massif and lunar. New field work in the Laramie complex by Scoates (1992) shows a positive correlation between modal plagioclase and deformation. This observation suggests that solid state anorthositic diapirs may have developed within the cumulate piles of somewhat less anorthositic magma chambers. Such a phenomenon may explain the extremely plagioclase-rich nature of some lunar anorthosites (98-99% plagioclase).

The second formidable problem facing the polybaric model for terrestrial anorthosites is the petrogenesis of massifs with average plagioclase composition in the range of An_{40-45}. An evolved, but broadly basaltic magma (e.g. HLCA - above) is a plausible parent magma for more calcic massifs, such as Harp Lake Emslie (1980). However, if the difference in average plagioclase composition were solely the result of the pressure effect on Ab-An partitioning, then the parent basalt of the more alkaline anorthosites would have ponded at the implausible depths of 50 to 70 km. One might suggest more extensive fractionation of the basaltic magma at depth. Perhaps, but not in the presence of plagioclase because Sr in the plagioclase from the more alkaline Norwegian anorthosites is 30 to 50 % higher than that in the more calcic Nain anorthosites; and Sr is highest in the most sodic plagioclase — that from the Labrieville Complex (Emslie, 1985). Perhaps, the connection is a little more subtle: if similar basaltic magmas pond at different depths, then the magma that ponds at the deepest level would undergo the most extensive fractionation prior to reaching saturation with plagioclase — at which point anorthosite formation begins. Such a mechanism might account not only for more sodic plagioclase, but also generally higher levels of K, Ti, and P in associated rocks.

In order to address this problem, we have examined the liquidus equilibria of a primitive monzonorite ("jotunite") from the Rogaland district of Norway. This rock (Tj) is part of the chill margin of the Bjerkreim-Sokndal (B-S) layered intrusion which is itself part of the massif. The composition of this rock has been proposed as that of the parent magma of the B-S intrusion Duchesne and Hertogen (1988) and of the associated swarm of dikes with composition ranging from monzonorite to monzonite (mangerite) to quartz monzonite Duchesne et al. (1989). More importantly, compositions such as Tj have been suggested to have also been parental to the anorthosites at depth (Duchesne (1984). Some results from Vander Auwera and Longhi (1994) are shown in Fig. 6. At 5 kbar the inferred crystallization sequence — plag $\rightarrow$ plag + ol $\rightarrow$ plag + ol + ilm $\rightarrow$ plag $\pm$ ol + ilm + opx — is identical to that observed in the most primitive cycles of the B-S intrusion, and the compositions of the olivine and plagioclase are similar to, albeit slightly more evolved than, the most primitive compositions observed in the intrusion (Nielson and Wilson, 1991) . Al_2O_3 concentrations in opx are also consistent with crystallization at $P \leq 5$ kbar. So the parent magma of the B-S intrusion was a slightly less evolved version of TJ. Plagioclase and orthopyroxene appear together on the liquidus of TJ from 10 to 13 kbar, where clinopyroxene joins them. So these pressures represent the logical choice for the genesis of TJ by melting or fractionation processes. This pressure range is also that predicted for the origin of the aluminous orthopyroxene megacrysts found within the anorthosites proper.

Fig. 7 from Longhi and Vander Auwera (1994), which shows the relations of mineral and liquid compositions, poses a dilemma, however. In Fig. 7 parent magma compositions are shown to project from the plagioclase component close to the plane of pyroxene compositions at 13 (Tj) and 11.5 (HLCA) kbar. There is a thermal maximum on the plag + opx + aug liquidus boundary where it crosses the plane of pyroxene compositions as indicated by the arrows showing the

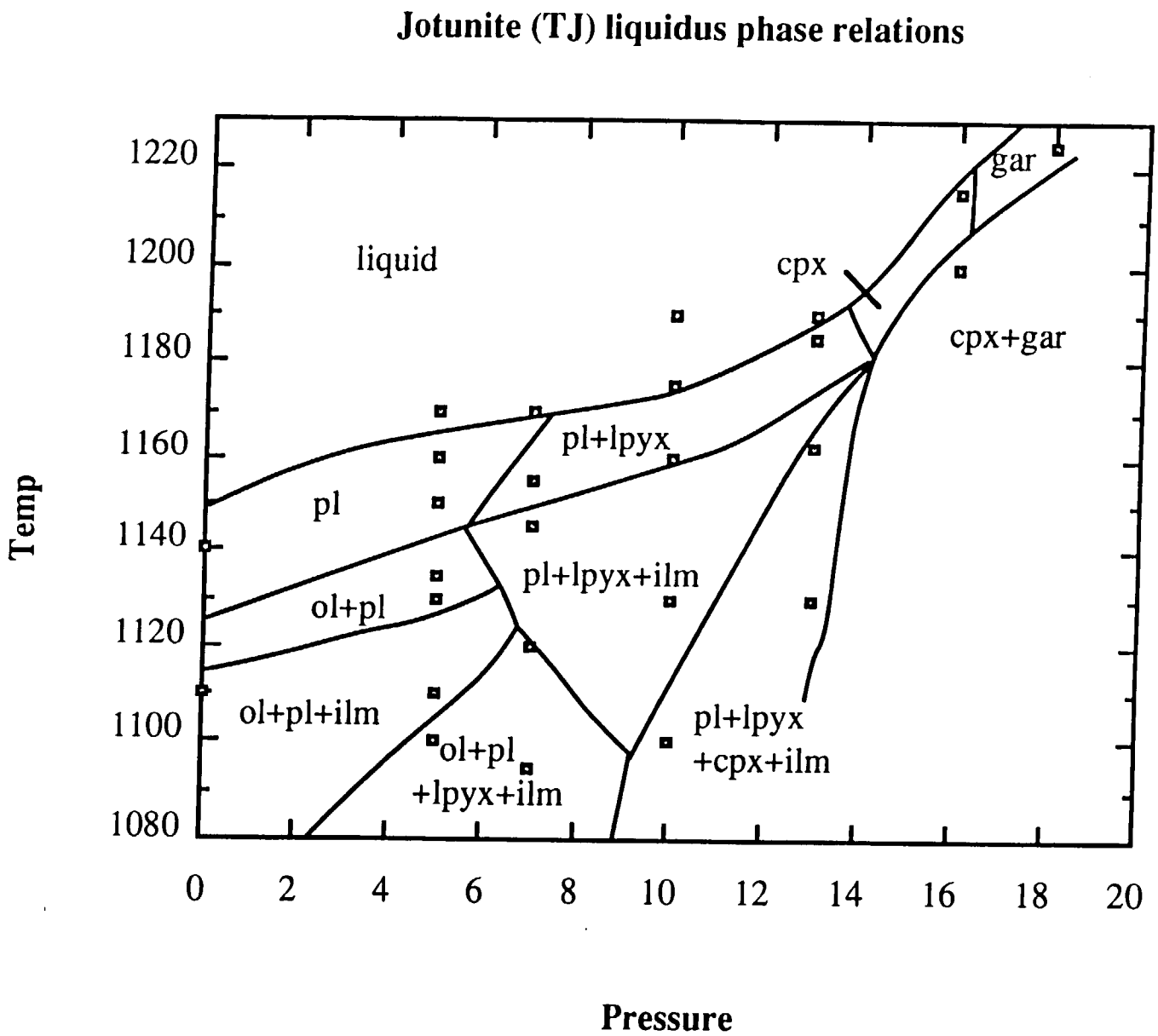

Fig. 6. Temperature (° C) - Pressure (kbar) melting relations of jotunite Tj (80,12,3A). Symbols: pl = plagioclase; ol = olivine; ilm = ilmenite; lpyx = low-Ca pyroxene; cpx = augite and/or pigeonite; gar = garnet. At 1 atm only experiments run at MW are shown.

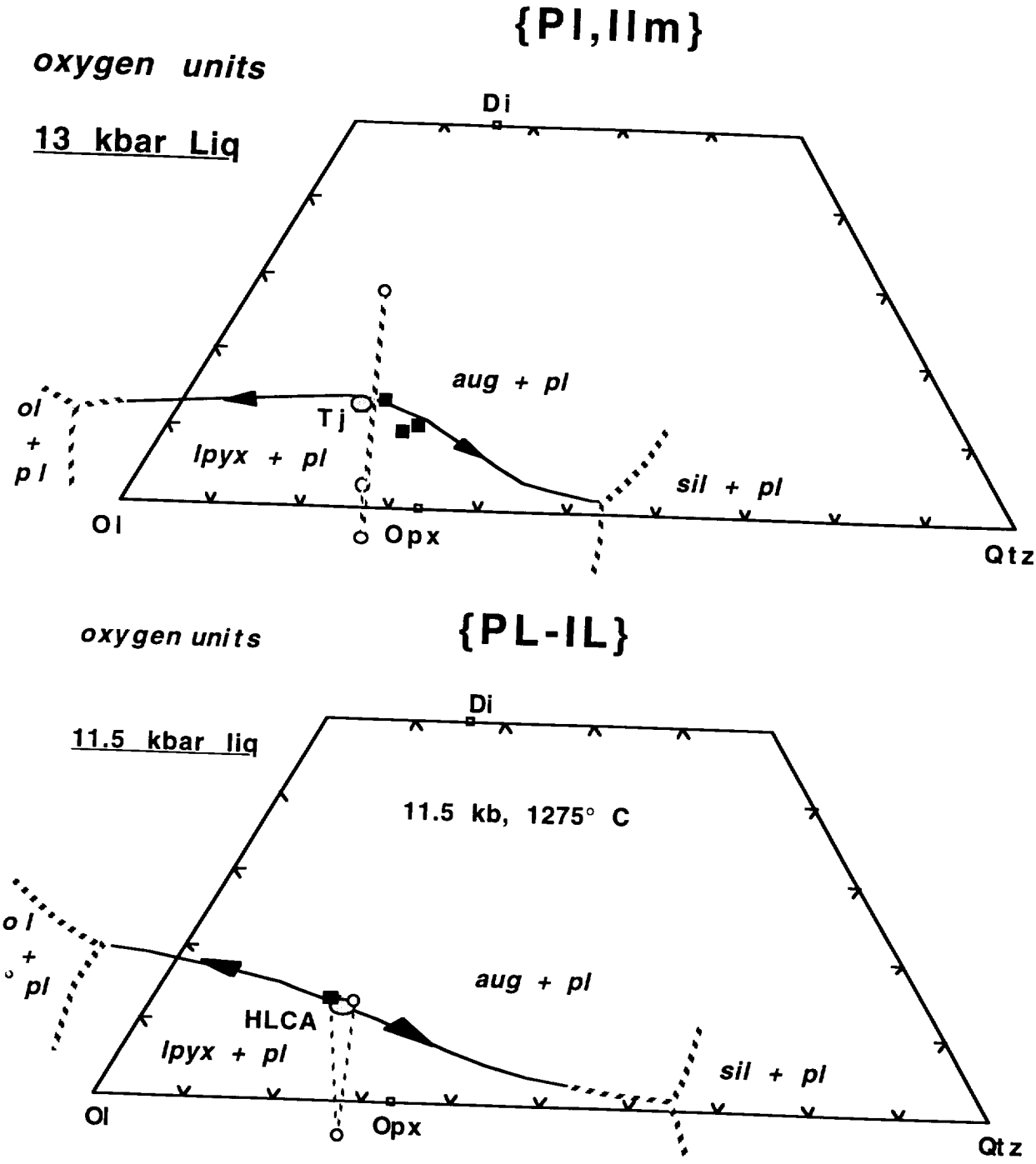

Fig.7 Projections from Plagioclase and Ilmenite components onto the Ol-Wo-Qtz plane. *Upper:* Solid squares are 13 kbar liquid compositions; dashed lines are tielines connecting coexisting pyroxene compositions (open circles); "lpyx" refers to low-Ca pyroxene, either orthopyroxene or pigeonite. Shaded oval is bulk composition of Tj. *Lower:* square is 11.5 kbar liquid composition. Shaded oval is bulk composition of HLCA.

direction of decreasing temperature. Adjacent to the thermal maximum are thermal ridges on the plag + opx and plag + aug liquidus surfaces. It should be clear, therefore, that in this pressure range TJ and similar compositions lie on the "thermal high-ground" — a peculiar situation for an evolved liquid. A similar situation exists for the more mafic Harp Lake parent, HLCA. Thus neither composition is likely to be reached by fractionation or even assimilation plus fractionation (AFC), because assimilation requires crystallization to provide the heat of melting and crystallization will drive the liquid composition away from the thermal divide. At lower pressures the pyroxene + plagioclase thermal divide disappears, so AFC processes can produce compositions similar to HLCA and Tj, but the range of primitive basalts is highly restricted because another thermal divide, ol+aug+plag, develops as soon as opx+cpx+plag disappears. The coincidence of AFC yielding significant amounts of liquids that both lie in thermal divides at higher pressures and are multi-saturated at the same higher pressures with crystals of the appropriate composition is rendered even less probable by the requirement for AFC to take place in the cooler middle crust. It thus appears that a variant of the lower crustal fusion hypothesis for the origin of the anorthosite parent magma is most consistent with the data.

REFERENCES

Delano, J. W. (1986) Pristine lunar glasses: criteria, data, and implications. *Proc. Lunar Planet. Sci. Conf. 16th, J. Geophys. Res., 91,* D201-D213.

Duchesne, J.C., and J. Hertogen (1988) Le magma parental du lopolithe de Bjerkreim-Sokndal (Norvège méridionale). *Comptes Rendus de l'Acadamie des Sciences Paris,* 306, 45-48.

Duchesne, J.C., E. Wilmart, D. Demaiffe, and J. Hertogen (1989) Monzonorites from Rogaland (Southwest Norway): a series of rocks coeval with but not comagmatic with massif-type anorthosites. *Precamb. Res., 7,* 111-128.

Duchesne, J.C. (1984) Massif anorthosites: another partisan review. In W.L. Brown, Ed., *Feldspars and feldspathoids,* p. 411-433, Reidel, Boston.

Emslie, R. F. (1970) The geology of the Michikamau intrusion, Labrador. *Geol. Surv. Can. Paper 68-57,* 85 pp.

Emslie, R. F. (1980) Geology and petrology of the Harp Lake Complex, central Labrador: An example of Elsonian magmatism, *Geol. Surv. Can. Bull. 293,* pp. 1-136.

Emslie, R.F. (1985) Proterozoic anorthosite massifs. In A.C. Tobi and J.L.R Touret, Eds., *The Deep Proterozoic Crust in the North Atlantic Provinces,* p. 39-60. Dordrecht, Netherlands.

Fram, M.S. and J. Longhi (1992) Phase equilibria of dikes associated with Proterozoic anorthosite complexes, *Am. Mineral., 77,* 605-616.

German, R.M. (1985) The contiguity of liquid phase sintered microstructures. *Metallurgical Transactions, 16A,* 1247-1252.

Gill, J. B. (1981) *Orogenic Andesites and Plate Tectonics.* Springer-Verlag, New York, 390 pp.

Hess, P. C., M. J. Rutherford, R. N. Guillemette, F. J. Ryerson, and H. A. Tuchfeld (1975) Residual products of fractional crystallization of lunar magmas: an experimental study, *Proc. Lunar Sci. Conf. 6th,* 895-910.

Klein, E.M. and C.H. Langmuir (1987) Global correlation of ocean ridge basalt chemistry with axial depth and crusta; thickness, *J. Geophys. Res. 92,* 8089-8115.

James, O.B., M. M. Lindstron, and M. K. Flohr (1988) Ferroan anorthosite from lunar breccia 64435: Implications for the origin and history of ferroan anorthosite, *Proc. Lunar Planet. Sci. Conf. 19th,* p. 219-243.

Longhi, J. (1990) Silicate liquid immiscibility in isothermal crystallization experiments, *Proc. Lunar Planet. Sci. Conf. 20th,* p. 13-24.

Longhi, J. (1991a) Comparative liquidus equilibria of hypersthene-normative basalts at low-pressure, *Amer. Mineral.,* 76, 785-800.

Longhi, J. (1991b) Complex magmatic processes on Mars: inferences from the SNC meteorites, *Proc. Lunar Planet. Sci. Conf. 21st,* p. 695-709.

Longhi, J. (1992a) Origin of green glass magmas by polybaric fractional fusion, *Proc. Lunar Planet Sci. 22,* 343-353.

Longhi, J. (1992b) Experimental petrology and petrogenesis of mare volcanics, *Geochim. Cosmochim. Acta, 56,* 2235-2251.

Longhi, J. (1992c) Volatiles in SNC petrogenesis: a Sr signal? *Lunar and Planetary Science XXIII,* The Lunar and Planetary Institute, pp. 805-806.

Longhi, J. (1993) Liquidus equilibria of lunar analogs at high pressure, *Lunar and Planetary Science XXIV,* The Lunar and Planetary Institute, pp. 895-896.

Longhi, J. (1994) Liquidus equilibria of lunar analogs in the garnet stability field, *Lunar and Planetary Science XXV,* The Lunar and Planetary Institute, pp. 803-804.

Longhi, J., and Ashwal, L.D. (1984) A two stage model for lunar anorthosites: An alternative to the magma ocean hypothesis, *Journal of Geophysical Research, 90,* C571-C584.

Longhi, J., E. Knittle, J. R. Holloway, and H. Wänke (1992) The bulk composition, mineralogy, and internal structure of Mars, in *MARS*, H. H. Kieffer, B. M. Jakosky, C. W. Snyder, and M. S. Matthews eds., pp. 184-208, The Univ. of Arizona Press, Tuscon.

Longhi, J., M.S. Fram, J. Vander Auwera, and J. Monteith (1993) Pressure effects, kinectics, and rheology of anorthositic and related magmas, *Am. Mineral., 78,* 1016-1030.

Longhi, J. and Vander Auwera, J. (1993) The monzonorite-anorthosite connection: the petrogenesis of terrestrial KREEP, *Lunar and Planetary Science XXIV,* The Lunar and Planetary Institute, pp. 897-898.

Longhi, J. and Vander Auwera, J. (1994) A mantle-derived parent magma for massif anorthosites??? In *Anorthosites, Rapakivi Granites and Related Rocks,* (R. F. Emslie, ed.) International Geological Correlation Programme Joint Meeting IGCP #290 and #315, p. 6.

Morse, S. A. (1982) A partisan review of Proterozoic anorthosites, *Amer. Mineral., 67,* pp. 1087-1100.

Nielsen, F.M. and J.R. Wilson (1991) Crystallization processes in the Bjerkreim-Sokndal layered intrusion, south Norway: evidence from the boundary between two macrocyclic units. *Contrib. Mineral. Petrol., 107,* 403-414.

Rutherford, M. J., P. C. Hess, and G. H. Daniel (1974) Experimental liquid line of descent and liquid immiscibility for basalt 70017, *Proc. Lunar Sci. Conf. 5th,* 569-583.

Scoates, J.S. (1992) Anorthosites in the Laramie Anorthosite Complex Wyoming, USA: structural and geochemical evidence for a complex emplacement, crystallization, and deformational history (abs.). In J. C. Duchesne, Ed., *Magma chambers and processes in anorthosite formation,* p. 21, International Geological Correlation Programme Project 290.

Spera, F.J. (1991) Lunar magma transport phenomena. *Geochim. Cosmochim. Acta., 56,* 2235-2251.

Taylor S. R. (1982) *Planetary Science: A Lunar Perspective*, The Lunar and Planetary Institute, Houston, 481 pp.

Vander Auwera, J. and J. LONGHI (1994) Experimental study of a jotunite: constraints on the parent magma composition and crystallization conditions (P, T, fO2) of the Bjerkreim-Sokndal layered intrusion (Norway). *Contrib. Mineral. Petrol.*, in press.

Warren, P. H., G. J. Taylor, K. Keil, D. N. Shirley, and J. T. Wasson (1983) Petrology and chemistry of two "large" granite clasts from the Moon, *Earth Planet. Sci. Lett., 64*, 175-185.

Warren, P. A., E. A. Jerde and G. W. Kallemeyn (1991) Prisitne moon rocks: Apollo 17 anorthosites. *Proc. Lunar Planet. Sci. Conf. 21st,* 51-61.

Wiebe, R. A. (1990) Evidence for unusually feldspathic liquids in the Nain Complex, Labrador, *Am. Mineral., 75,* 1-12.

APPENDED PUBLICATIONS

Short Abstracts

1. FRAM, M.S. and J. LONGHI (1989) Dike phase equilibria and the origin of Proterozoic Anorthosites, *EOS*, p. 1395.

2. LONGHI, J. (1990) Dynamical melting models for the eucrite parent body, *EOS, 71,* p. 1433.

3. FRAM, M.S. and J. LONGHI (1991) Plagioclase/melt partitioning as a function of pressure, *EOS, 72*, p. 304-305.

4. LONGHI, J. (1991) A comparison of incompatible trace element patterns in terrestrial, lunar, and martian basalts, *EOS, 72,* p.281.

5. LONGHI, J. and J. VANDER AUWERA (1992) Polybaric fractionation and the connection between high-Al gabbro and monzonorite, *Magma Chambers and Processes in Anorthosite Formation,* International Geological Correlation Project 290.

6. VANDER AUWERA, J. and J. LONGHI (1992) Phase equilibria and fractionation paths from 1 atm up to 16 kb of the monzonorite Tjorn chilled facies: Implications for the Bjerkreim-Sokndal lopolith (southern Norway), *Magma Chambers and Processes in Anorthosite Formation,* International Geological Correlation Project 290.

7. LONGHI, J. (1992) Phase equilibria and crystal-liquid partitioning for lunar picritic compositions, *EOS, 73,* p. 615.

8. LONGHI, J. and R. J. KINZLER (1993) Supercritical low-Ca clinopyroxene on the solidus in the garnet stability field, *EOS, 74,* p. 658.

9. VANDER AUWERA, J., J. LONGHI, and J.-C. DUCHESNE (1993) Jotunites from the Rogaland Province (Norway): constraints from experimental data and the partitioning of Sr (plag/melt) and Cr (opx/melt), *EOS, 74,* p. 659.

10. LONGHI, J. and VANDER AUWERA, J. (1994) A mantle-derived parent magma for massif anorthosites??? In *Anorthosites, Rapakivi Granites and Related Rocks,* (R. F. Emslie, ed.) International Geological Correlation Programme Joint Meeting IGCP #290 and #315, p. 6.

Extended Abstracts and Short Papers

11. LONGHI, J. (1989) Differentiates of the picritic glass magmas: the missing mare basalts, *Workshop on Lunar Volcanic Glasses: Scientific and Resource Potential,* The Lunar and Planetary Institute, pp. 15-16.

12. LONGHI, J. (1989) Complex magmatic processes on Mars, *Workshop on Evolution of Magma Bodies on Mars,* The Lunar and Planetary Institute, pp. 25-26.

13. LONGHI, J. (1990) Magmatic processes on Mars: insights from SNC meteorites, *Lunar and Planetary Science XXI*, The Lunar and Planetary Institute, pp. 716-717.

14. FRAM, M.S. and LONGHI, J. (1990) Experimental petrology of dikes related to anorthosite complexes, *Lunar and Planetary Science XXI*, The Lunar and Planetary Institute, pp. 385-386.

15. LONGHI, J. (1990) Dynamical melting models of mare basalts, *Workshop on Mare Volcanism and basalt Petrogenesis,* The Lunar and Planetary Institute, pp. 15-16.

16. LONGHI, J. and V. PAN (1991) Implications of the Venera 13 surface analysis for the CO_2 content of the interior of Venus, *Lunar and Planetary Science XXII,* The Lunar and Planetary Institute, pp. 829-830.

17. LONGHI, J. (1991) The origin of the green glass magmas by polybaric partial melting, *Lunar and Planetary Science XXII,* The Lunar and Planetary Institute, pp. 831-832.

18. LONGHI, J., FRAM, M. S., VANDER AUWERA, J., and MONTIETH, J. (1992) Pressure effects in anorthositic and related magmas, *Lunar and Planetary Science XXIII,* The Lunar and Planetary Institute, pp. 803-804.

19. LONGHI, J. (1992) Volatiles in SNC petrogenesis: a Sr signal? *Lunar and Planetary Science XXIII,* The Lunar and Planetary Institute, pp. 805-806.

20. LONGHI, J. (1992) Conjuring up the Red Planet, *Lamont-Doherty Geological Observatory 1990 & 1991 Report,* pp. 98-104.

21. LONGHI, J. (1993) Liquidus equilibria of lunar analogs at high pressure, *Lunar and Planetary Science XXIV,* The Lunar and Planetary Institute, pp. 895-896.

22. LONGHI, J. and VANDER AUWERA, J. (1993) The monzonorite-anorthosite connection: the petrogenesis of terrestrial KREEP, *Lunar and Planetary Science XXIV,* The Lunar and Planetary Institute, pp. 897-898.

23. LONGHI, J. (1994) Liquidus equilibria of lunar analogs in the garnet stability field, *Lunar and Planetary Science XXV,* The Lunar and Planetary Institute, pp. 803-804.

Journal Articles (abstracts only)

24. LONGHI, J. (1990) Silicate liquid immiscibility in isothermal crystallization experiments, *Proc. Lunar Planet. Sci. Conf. 20th*, p. 13-24.

25. LONGHI, J. (1991) Complex magmatic processes on Mars: inferences from the SNC meteorites, *Proc. Lunar Planet. Sci. Conf. 21st,* p. 695-709.

26. LONGHI, J. (1991) Comparative liquidus equilibria of hypersthene-normative basalts at low-pressure, *Amer. Mineral.*, 76, 785-800.

27. LONGHI, J. (1992) Origin of green glass magmas by polybaric fractional fusion, *Proc. Lunar Planet Sci. 22,* 343-353.

28. FRAM, M.S. and J. LONGHI (1992) Phase equilibria of dikes associated with Proterozoic anorthosite complexes, *Am. Mineral., 77,* 605-616.

29. LONGHI, J. (1992) Experimental petrology and petrogenesis of mare volcanics, *Geochim. Cosmochim. Acta, 56,* 2235-2251.

30. LONGHI, J., E. KNITTLE, J. R. HOLLOWAY, and H. WÄNKE (1992) The bulk composition, mineralogy, and internal structure of Mars, in *MARS*, H. H. Kieffer, B. M. Jakosky, C. W. Snyder, and M. S. Matthews eds., pp. 184-208, The Univ. of Arizona Press, Tuscon.

31. LONGHI, J., M.S. FRAM, J. VANDER AUWERA, and J. MONTEITH (1993) Pressure effects, kinectics, and rheology of anorthositic and related magmas, *Am. Mineral., 78,* 1016-1030.

32. BEATTIE, P.D., M.J. DRAKE, J.H. JONES, W.P. LEEMAN, J. LONGHI, G.A. MCKAY, R.H. NIELSEN, H. PALME, D.M. SHAW, E. TAKAHASHI, and E.B. WATSON (1993) A terminology for trace element partitioning. *Geochimica Cosmochimica Acta, 57,* 1605-1606.

33. VANDER AUWERA, J. and J. LONGHI (1994) Experimental study of a jotunite: constraints on the parent magma composition and crystallization conditions (P, T, fO2) of the Bjerkreim-Sokndal layered intrusion (Norway). *Contrib. Mineral. Petrol.,* in press.

Dike Phase Equilibria and the Origin of Proterozoic Anorthosites

Miranda S Fram and John Longhi (Lamont-Doherty Geological Observatory of Columbia University, Palisades, NY 10964)

The composition of magmas parental to Proterozoic anorthosites has eluded petrologists for many years. Proposed compositions range from anorthositic to basaltic. Dikes found in anorthosite complexes may represent possible parental magmas. The Nain complex in Labrador contains anorthositic dikes, described by Weibe (1979), which have similar bulk composition to the anorthosite plutons in the area. The Harp Lake Complex, also in Labrador and described by Emslie (1980), has dikes and gabbros marginal to the anorthosite plutons which are high-Al basalts.

We conducted 1 atm and piston cylinder experiments up to 20 kb on a natural sample from the Nain dikes (500B) and a synthetic sample representing an average of the Harp Lake high-Al dikes and gabbros (HLC-A). Sample 500B crystallizes a large amount of plagioclase (An47-60) between 1 atm and 20 kb, but the liquidus temperatures over that range, 1365° to 1410°, seem unreasonably high for 500B to ever have been wholly liquid. The second phase to appear at all pressures is augite, whereas the dike is dominated by pigeonite, which suggests 500B is a cumulate rather than a liquid with only accumulated plagioclase. Sample HLC-A crystallizes a substantial volume of plagioclase (An50-60) alone at pressures between 1 atm and 10 kb before reaching saturation with a second phase which is olivine at low pressure and low-Ca pyroxene at high pressure. At 15 kb, low-Ca pyroxene replaces plagioclase as the liquidus phase. Calculations indicate that adding 30% low-Ca pyroxene and 10% olivine to HLC-A generates a composition similar to basalts equilibrated with peridotite at 15 kb in experiments by Takahashi and Kushiro (1983). In preliminary experiments on a synthetic analogue of the basalts, low-Ca pyroxene forms at 15 kb.

Our results support a polybaric fractionation model for the production of parental magmas of Proterozoic anorthosites. Mantle melts crystallize mafic minerals at depth and then as the magmas rise, the phase boundaries shift away from the plagioclase component with falling pressure, causing the magmas to become oversaturated with plagioclase.

1. 1989 Fall Meeting

2. AGU # 008085677

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(b) 914-359-2900 x661

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Dynamical Melting Models for the Eucrite Parent Body

J Longhi (Lamont-Doherty Geological Observatory, Palisades, NY 10964; 914-359-2900)

A variety of petrogenetic schemes have been proposed for eucrites including isobaric partial melting and polybaric fractional crystallization. This paper examines some of the consequences of polybaric porous flow melting of small parent bodies with compositions similar to those proposed for the eucrite parent body (EPB). A regime of porous flow would presumably exist after modest amounts of external and radiogenic heating or after rapid solidification of a magma ocean (no insulating crust). True porous flow melting was bracketted by two endmember models: one was partial fractional fusion in which a fixed amount of equilibrium melt was removed from the decompressing source at each pressure step and allowed to accumulate; the other was progressive pressure-release melting in which the source maintained its composition despite extraction and accumulation of melt. These two idealized models produce significantly different compositions of accumulated melt only when operating over 1 or more GPa, so true porous flow may be approximated by either given the low-pressures likely on the EPB. For a variety of initial starting pressures ($\leq$ 0.5 GPa) only the more ferroan source compositions, i.e. Mg/(Mg+Fe) $\leq$ 0.7, can produce eucritic melts. The pressure that would be inferred from the composition of the accumulated melt is similar to the average pressure of the flow regime, so the proximity of the eucrites to 1 atm cotectics suggests a flow regime corresponding to $\leq$ 0.3 GPa. Relatively high porosities (~10%) are required to produce sufficient melting to exhaust plagioclase in the source and eliminate the negative Eu-anomaly. Porous melting over a wider pressure interval (0.5 GPa) or starting with a less aluminous source can produce diogenitic magmas. Relatively high degrees of melting ($\geq$ 20%) for the EPB inferred from siderophile element fractionations in eucrites together with the inferences from this study are consistent with a multi-stage evolution for the EPB: an initial melting event (magma ocean?) during which core formation and separation of olivine occured at depth followed by a subsequent porous flow regime in the more ferroan outer residuum; following the crystallization of the eucrites, the parent body (perhaps as large as 600 km in radius) was fragmented by impacts.

1. 1990 Fall Meeting

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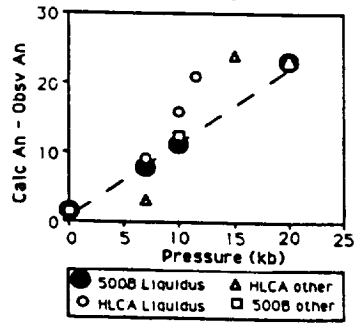
Plagioclase/Melt Partitioning as a Function of Pressure

M S Fram and J Longhi (Lamont-Doherty Geological Observatory of Columbia University, Palisades, NY 10964)

We conducted melting experiments from 1 bar to 30 kb on two dike compositions associated with Proterozoic Massif anorthosites. One composition (HLCA - high aluminum gabbro) has plagioclase on the liquidus to 11.5 kb; the other (500B - anorthosite) has plagioclase on the liquidus to >20 kb. Along the liquidus of each composition there is a systematic shift in the plagioclase composition with increasing pressure. For example, in 500B the liquidus plagioclase changes from An_{80} at 1 bar, 1365°C to An_{55} at 20 kb, 1420°C. These shifts occur at relatively constant liquid composition. In order to separate the effect of increasing temperature from pressure we compare the observed plagioclase composition to the predicted composition at 1 bar and experimental run temperature using the model of Drake (1976, *GCA, 40*) There is a marked divergence between the observed and the calculated plagioclase with increasing pressure that is approximately linear. For 500B the slope is ~1.12 An units per kbar. The strong pressure dependence of plagioclase melt partitioning has consequences for models of massif anorthosites, which typically



have intermediate plagioclase. A number of studies have suggested that the parent magma for the anorthosites is basaltic. The present results show that basaltic liquids that crystallize calcic plagioclase at low pressure will crystallize more intermediate plagioclase at lower crustal pressures. Transport of this intermediate plagioclase as crystal-rich suspensions to upper crustal sites of intrusion may account for the calculated low pressures of intrusion and the intermediate plagioclase compositions of massif anorthosites.

A Comparison of Incompatible Element Patterns in Terrestrial, Lunar, and Meteoritic Basalts

J Longhi (Lamont-Doherty Geological Observatory, Palisades, NY 10964; 914-359-2900x659)

Each planetary object for which we have samples has a unique magmatic style. These styles may be distinguished by comparison of chondrite-normalized patterns of refractory incompatible lithophile elements and isotope systematics. It is useful to consider the patterns formed when these elements are plotted in order of apparent incompatibility and to discuss these patterns in terms of highly incompatible (HI = Ba, Th, Ta, Nb, light REE), moderately incompatible (MI = Hf, Zr, middle REE), and weakly incompatible (WI = Ti, heavy REE) elements. Samples of the Eucrite Parent Body (EPB), an asteroid, show the simplest patterns (nearly flat and featureless), reflecting a relatively simple igneous history. The Earth, of course, has the widest variety of magmatic styles and hence incompatible element patterns; but the most abundant basalts (mid ocean ridge, ocean island, flood) and also komatiites have relatively featureless patterns that vary mainly in terms of relative enrichment or depletion of the HI elements. Isotope analyses show that nearly all of these basalts come from sources with long term depletions of the HI elements, so that HI element enrichments must be due to some recent igneous process, probably small degrees of melting coupled with porous flow. Overall HI element enrichments also correlate with local enrichments of Nb (≈Ta) and Ti relative to the HI and WI, respectively. Island arc basalts provide the major exceptions to the terrestrial patterns with dramatic depletions of Nb (≈Ta) relative to the HI and of Zr(≈Hf) relative to the MI. By way of contrast, the incompatible element patterns of most lunar basalts are feature-rich with the depletion of Eu relative to the MI elements, which reflects the mobility of plagioclase during early lunar differentiation, being the most famous. However, there are also prominent enrichments of Ta(≈Nb) and Ti that correlate with Ti concentration in the mare basalts; whereas Ta(≈Nb) and Ti show prominent depletions in KREEP basalts. These patterns reflect the mobility of ilmenite during early lunar differentiation -- even in the source regions of the low-Ti basalts -- and also the generally complementary relation of KREEP and the mare basalt source region. Martian samples, the shergottite-nakhlite-chassignite (SNC) meteorites, have both prominent enrichments and depletions of Ta(≈Nb) and Zr(≈Hf) and lack Eu-anomalies. The Nakhlites show strong overall HI enrichments, yet come from very strongly depleted source regions, suggesting miniscule degrees of partial melting -- much like terrestrial ocean island basalts. However, nakhlites also strong relative depletions of Ta(≈Nb) and Zr(≈Hf) -- like arc basalts. Shergottites show overall depletions of the HI elements, but prominent relative enrichments of Ta(≈Nb) and Zr(≈Hf). In this regard, they are complementary to nakhlites and unlike any other common solar system basalts.

An important question to answer is whether the process (still controversial) that produces relative depletions of Nb (≈Ta) and Zr(≈Hf) in arc basalts is also operative on Mars in the absence of plate tectonics. There is a subtle distinction between the terrestrial and martian patterns, however, that may indicate separate processes: in terrestrial basalts the magnitude of the Nb(≈Ta) anomaly, whether negative (arcs) or positive (ocean islands), is always larger than the Zr(≈Hf) anomaly; whereas in the martian patterns the Ta(≈Nb) and Zr(≈Hf) anomalies have similar magnitudes. At the very least, the martian patterns indicate that Mars' igneous evolution is not simply a mixture of lunar and terrestrial styles.

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5. (a) P03
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POLYBARIC FRACTIONATION AND THE CONNECTION BETWEEN HIGH-AL GABBRO AND MONZONORITE

John Longhi (Lamont-Doherty Geological Observatory, Palisades, NY 10964, USA) and Jacqueline Vander Auwera (LA Géologie, Pétrologie, Géochimie - B20- Université de Liège, B-4000 Sart Tilman, Belgium)

There is a near continuum in major and minor element composition between high-Al gabbros and ferrodiorites from the Harp Lake Complex of Labrador [1]. This continuum is overlapped at intermediate MgO concentrations and extended to lower MgO and higher SiO2 by monzonoritic-syenitic dikes from the Rogaland Complex of Norway [2]. When compared to decreasing MgO (i.e., progressive differentiation), Al2O3 decreases, SiO2 initially remains constant and then both increases and decreases, P2O5 shows a weak overall enrichment in the Harp Lake suite, but a precipitous increase followed by a sharp decrease in the Rogaland suite, TiO2 and FeO first increase then decrease, and K2O increases. The compositions of liquids produced in melting experiments on two starting materials, a synthetic high-Al gabbro from Harp Lake (HCLA) and rock powder from the fine-grained monzonoritic marginal facies (TJ) of the Bjerkreim-Sokndal Lopolith, generally fall along limited portions of these trends and support a connection between a gabbroic parent and monzonoritic derivatives. The connection, however, cannot be simple.

Only one element's variation seems straightforward: the sudden decrease in TiO2 beginning at about 4 wt% MgO is due to the precipitation of ilmenite. With decreasing MgO, SiO2 in the HCLA liquids remains approximately constant, but the TJ liquids produced at high pressure in graphite capsules show a weak decrease in SiO2, whereas those run at 1 bar at NNO show an increase. This difference in behavior is due to the early appearance of magnetite in the 1 bar experiments, so differences in oxidation state may play a role in determining fractionation paths. There is insufficient K2O in the gabbro to account for the K2O in the monzonorites by simple fractional crystallization. So, as is often done in the case of convergent margin volcanics (e.g.,[3]), it is necessary to invoke crustal contamination. Among Rogaland samples [2], Sr isotopes also show crustal affinities but lack of correlation between I_{Sr} and K, Rb suggests that if contamination occurs, it cannot be simple. Fractionation, even coupled with crustal assimilation, is unlikely to explain the sharp drop in SiO2 and the sharp increase in P2O5 observed in the natural samples. Flow differentiation within the monzonoritic dikes and migration of the interstitial liquid with respect to a crystalline matrix of low-Ca pyroxene, ilmenite, and apatite seem to be the most likely explanations of the low-SiO2/high-P2O5 trend.

These observations support a complex connection between high-Al gabbros and monzonorites: fractional crystallization plus crustal assimilation at 8 to 12 kb links the gabbroic magma to the more primitive monzonorites; higher level differentiation produces the variations in the more evolved monzonorites and syenites. This connection is also consistent with a range of parent magmas from high-Al gabbro to monzonorite producing a range of massif-type anorthosites.

REFERENCES

[1] Emslie, RF (1980) Geol Surv Can Bull 293: 1-136; [2] Duchesne JC, Wilmart E, Demaiffe D, Hertogen J (1989) Precambrian Research 45: 111-128; [3] Grove TL, Gerlach DC, Sando TW (1982) Contrib Mineral Petrol 80:160-182.

PHASE EQUILIBRIA AND FRACTIONATION PATHS FROM 1 ATM UP TO 16 KB OF THE MONZONORITIC TJORN CHILLED FACIES : IMPLICATIONS FOR THE BJERKREIM-SOKNDAL LOPOLITH (SOUTHERN NORWAY)

J VANDER AUWERA (LA Géologie-Pétrologie-Géochimie (B20), Université de Liège, Belgium) and J LONGHI (Lamont-Doherty Geological Observatory, Palisades, NY 10964, USA)

A series of experiments were performed from 1 atm up to 16 kb on the so-called Tjörn chilled facies of the Bjerkreim-Sokndal lopolith in order to confirm or infirm its suitability as a parent magma of this layered intrusion [1]. These experimental data also put constraints on the still debated pressure of emplacement of the intrusion. [2] and [3] proposed a pressure of 3-4 kb on the basis of the osumilite occurrence in the contact aureole whereas [4] argued that orthopyroxene-fayalite-quartz assemblages in the upper part of the Bjerkreim-Sokndal lopolith give a pressure of 6-7 kb. According to [5], CO2-rich fluid inclusions occurring in the charnockitic upper part of the lopolith suggest a high pressure of 7.4 + 1 kb.

One-atmosphere experiments were run in a Deltech gas mixing furnace (CO-CO2 atmosphere) at fO_2 conditions of the NNO buffer as suggested by ilmenite-magnetite equilibria in the lopolith [6]. High pressure experiments (5-7-10-13-16 kb) were run in standard [7] piston cylinders using graphite capsules and barium carbonate pressure cells. fO_2 was grossly approximated to be below QFM. Plagioclase is the liquidus phase from 1 atm (An_{60}) up to 7kb (An_{48}) followed by olivine at low pressure (1 atm up to 5 kb) or at 7 kb. From 10 kb to 13 kb, both plagioclase and orthopyroxene are the liquidus phases, followed by augite, pigeonite and ilmenite at lower temperature. At 16 kb garnet and augite are the liquidus phases. The effect of the relatively lower oxidation state superimposed by the graphite capsules is to reduce the amount of ferric iron and increase the fayalite component of the starting material. As a result, the Mg# of olivine and pyroxene is somewhat lower than it should be and the stability of olivine relative to low-Ca pyroxene is enhanced. As a result, comparison of experimental and natural plagioclase composition is a more reliable test of Tjörn as a parent magma, and a limit of olivine stability at high pressure in the experiments is an upper limit for pressure of crystallization of the intrusion. The least evolved cumulate in the lopolith contains olivine (Fo73), plagioclase (An50), and ilmenite. The plagioclase composition at the liquidus of Tjörn at 7Kb is An48, which agrees fairly well. However, orthopyroxene not olivine is the near-liquidus mafic silicate at 7kb. Olivine and ilmenite are the near-liquidus mafic minerals at 5 kb, so the pressure of crystallization for the intrusion is 5 kb or less. Experimental data [8] show that the anorthite content of the liquidus plagioclase will increase with decreasing pressure by about 1% per kb, so a pressure of 5kb would yield an exact match between experimental and natural plagioclase. Thus the Tjörn chilled facies sample appears to be a good candidate for the parent magma of cycles MCU III and IV in the Bk-Sk lopolith.

Liquid compositions obtained at different pressures show consistent fractionation trends in the range 5% to 3 % MgO : TiO2 first increases and then sharply decreases due to ilmenite crystallization; P2O5, K2O and FeO_t strongly increase; Na2O as well as SiO2 and Al2O3 slightly decrease due to plagioclase crystallization; CaO is almost constant. Fractionation trends of monzonoritic dykes [9] display similar features except for P2O5 and TiO2 which sharply decrease. Due to higher P2O5 content in these monzonorites, apatite is a liquidus phase. This is also the case for ilmenite despite similar TiO2 contents in Tjörn and monzonorites.

REFERENCES

[1] Duchesne JC and Hertogen J (1988) C R Acad Sci Paris 306(II):45-48. [2] Maijer C, Andriessen PAM, Hebeda EH, Jansen JBH and Verschure RH (1981) Geologie en Mijbouw 60: 267-272. [3] Jansen JBH, Scheelings M and Bos A (1985) NATO Advanced Study Institute. Series C158 : 499-516. [4] Wilmart E and Duchesne JC (1987) Norsk Geologisk Tidsskrift 67:185-196. [5] Wilmart E, Clocchiatti R, Duchesne JC, Touret JLR (1991). Contrib Mineral Petrol 108:453-462. [6] Duchesne JC (1972) J Petrol 13:57-81. [7] Boyd JR and England JL (1960) J Geophys Res 65 : 741-748. [8] Longhi J, Fram M, Vander Auwera J, Montieth J (1992) Lunar and Planetary Science Conf Abstract. [9] Duchesne JC, Wilmart E, Demaiffe D, Hertogen J (1989) Precamb Res 45:111-128.

Phase Equilibria and Crystal-Liquid Partitioning for Lunar Picritic Compositions

J Longhi (Lamont-Doherty Geological Observatory, Palisades NY 10964; 914-359-2900 x659)

Melting experiments on synthetic low-alkali (lunar) picritic compositions in iron capsules at 25 to 35 kb illustrate many of the predicted effects of pressure and composition on liquidus boundaries and partition coefficients. For example, in liquids with Mg' [MgO/(MgO+FeO)] ~ 0.5 and saturated with olivine (ol) and orthopyroxene (opx) the normative olivine content is higher than in similarly saturated melts with Mg' = 0.7-0.8. At 25 kb the compositions of ol+opx-saturated and ol+cpx+gar-saturated liquids require that opx, but not ol, is in reaction with lherzolitic liquids (i.e., liq + opx = ol + cpx + gar). This means that the amount of opx will increase with melting until cpx or gar is exhausted.

The simple molar partition coefficients (D*) for Al_2O_3 between opx-liq (0.4-0.5) and cpx-liq (~0.6) at 25 to 30 kb are approximately 4 times their 1 atm values, whereas D* for TiO_2 for both opx-liq and cpx-liq (~0.1) does not change with pressure. Likewise, K_D values for olivine (0.30-0.35) are close to 1 atm values (0.331 ± .014) in lunar basalts. K_D values for opx (0.28-0.31) and cpx (0.30) are slightly lower than ol values; but K_D for gar (0.44) is distinctly higher. Cr_2O_3 is highly compatible in gar (D* = 2.6-3.3), whereas TiO_2 is incompatible (D*=0.22-30).

These preliminary results are in accord with estimates of the composition and pressure dependences of liquidus boundaries made by Longhi (1992) and support the derivation of picritic lunar magmas by polybaric fractional fusion.

Ref: Longhi (1992) *Proc. Lunar Planet. Sci. 22*, 343-353.

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2. 000066816
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   (c) 914-365-3183
4. V
5. (a)
   (b) 3630 experimental petrol.
   3640 igneous petrology
6. Oral
7. 0%
8. $50 charge to Am. Express
   [redacted]
   John Longhi exp 12/92
9. C
10.
11. No

Supercritical low-Ca Clinopyroxene on the Solidus in the Garnet Stability Field

John Longhi and Rosamond J. Kinzler (Lamont-Doherty Earth Obs., Palisades, NY 10964; 914-365-8659;
e-mail: longhi@lamont.ldeo.columbia.edu)

High-pressure melting experiments (graphite capsules) on anhydrous glass + crystal mixtures with modified mantle compositions show that a low-Ca clinopyroxene coexists with garnet, orthopyroxene, and olivine (fo_{90}) on the depleted mantle solidus at 28 kbar. Bracketting runs constrain the clinopyroxene composition at < 9.0 wt% CaO and ~ 7 wt% Al_2O_3. These values are less than those observed in clinopyroxene (11, 9) coexisting with spinel, orthopyroxene, olivine, and melt at 23 kbar and suggest that the CaO content of clinopyroxene decreases with increasing pressure along the mantle solidus. Also, clinopyroxene compositions are apparently continuous in the range of 9 to 15 % CaO along the ol + gar + cpx liquidus curve. These observations are consistent with the predictions of Bertka and Holloway [*JGR*, in press] that the temperature of the terrestrial mantle solidus would begin to exceed that of the clinopyroxene solvus at ~25 kbar yielding a supercritical low-Ca clinopyroxene. Melting experiments on martian mantle analogs at 23 kbar and lunar compositions at 25 and 30 kbar also show a low-Ca clinopyroxene coexisting with orthopyroxene, garnet and olivine (fo_{75}), so a supercritical clinopyroxene is apparently a general feature of anhydrous mantle assemblages at high pressure.

Phase compositions constrain the melting reaction at 28 kbar to be ol + cpx + gar → opx + liq, as is also the case for the same, but less magnesian, lunar assemblage. This means that the proportion of orthopyroxene will increase with progressive melting of peridotite. The array of lunar phase compositions at 25 to 30 kbar also requires the relation ol + cpx → opx + liq along the garnet-free ol + opx + cpx liquidus curve. The same relation appears to hold at 28 kbar for the terrestrial compositions that we have examined.

The increasing solubility limits of MgO and Al_2O_3 in clinopyroxene solid solution are such that at higher pressures orthopyroxene and/or garnet may dissolve into clinopyroxene at the solidus, depending on the mantle composition.

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4. V
5. (a) None

   (b) 3630 Exp. mineralogy and petrology
   3640 Igneous petrology
   3615 Composition of the m
6. (a) Oral
7. 0%
8. $60 American Express John Longhi [redacted]
9. C
10. None; yes
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11. No

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- = 0.13, Nb = 0.006,
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, Nd = 0.04, Sm =

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V32A-8 1330h POSTER

Jotunites from the Rogaland Province (Norway) : constraints from experimental data and the partitioning of Sr (plag/melt) and Cr (opx/melt)

J Vander Auwera (Université de Liège - 4000 Sart Tilman - Belgium)
J Longhi (Lamont-Doherty Earth Observatory, Palisades, NY 10964)
JC Duchesne (Université de Liège - 4000 Sart Tilman - Belgium)

The DSr (plag/melt) and the DCr (opx/melt) have been measured with a CAMEBAX electron nucroprobe in melting experiments performed on the jotunitic border facies of the Bjerkreim-Sokndal lopolith (TJ) (Vander Auwera and Longhi, Terra Abs 1993). There is no significant pressure effect on the DSr from 1 atm up to 13 kb and depending on the anorthite content, DSr ranges from 1.7 (An60) to 1.9 (An43). These values are lower than those (2.5 (An60) to 3.6 (An43)) calculated with the regression of Blundy and Wood (GCA, 1991). DCr is lower than 3 at 1 atm (NNO buffer) and equals to 14 at 10 kb (< MW buffer). This last value is a minimum as DCr strongly increases with fO_2 (Barnes, GCA 1986).

Melting experiments have been run at 5 kb on the chilled margin and an FTP rock of the Varberg jotunitic dyke (Duchesne et al, CMP 1985). Comparison of phase equilibria in experiments and dyke suggests that the FTP rock is a crystal ladden liquid or even a cumulate. A least square regression model corroborates this conclusion. By eliminating these rocks from the Rogaland data set (Duchesne et al,1985) and using experimental data on TJ, the jotunitic liquid line of descent shows a first increase in FeO_t together with a decrease in SiO_2 followed by a strong decrease in FeO_t and an increase in SiO_2.

Major and minor element compositions suggest a connection between TJ and high-Al gabbros (HAG) through polybaric fractional crystallization coupled with crustal assimilation (Longhi and Vander Auwera, Lun Planet Sci XXIV, 1993). Apparently, this connection is not supported by trace element as some jotunites of Rogaland display no negative Eu anomaly and only a slight depletion in Sr relative to LREE. Nevertheless, calculation of the liquid REE patterns show that when plagioclase crystallizes cotectically with pyroxenes as observed in a high-Al gabbro at 11.5 kb (Fram and Longhi, Am Min 1992), a negative Eu anomaly appears only after 60% crystallization of HAG and is important only after 80% crystallization . Moreover, using the measured DSr and the same cumulate, the Sr depletion relative to LREE at 60% crystallization is similar to that observed in the least differentiated jotunites of Rogaland. These results give then further support to the connection between high-Al gabbros and jotunites. On the other hand, considering that the DCr measured at 10 kb is a minimum value, it is possible to crystallize Cr-rich opx from evolved liquids like TJ (Cr ≈ 50 ppm). This is consistent with parent magmas ranging from high-Al gabbros to jotunites in order to produce different massif type anorthosites.

V32B CA: Brooks Wed 1330h
Precambrian Magmatism Posters
Presiding: M A Hamilton, Geological Survey of Canada

V32B-1 1330h POSTER

Behavior of Hf and Nd isotopic systems in crustally-derived granitoids

P J Patchett and J D Vervoort (both at Dept. of Geosciences, University of Arizona, Tucson, AZ 85721)

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blueschist facies metamorphi
of Middle Jurassic - Early
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The data indicate that a portion of th
is at least as old as Proterozoic. Ou
Late Proterozoic U-Pb zircon ages ha
Brooks Range (Karl and Aleinikoff, 19
the Yukon-Tanana terrane (Aleinikoff et
present as detrital grains in metasedim
Grand Forks-Kettle complexes in southe
Washington State (Ross and Parrish, 199
magmatism is cryptic, however, it may be
rifting of North America.

V32B-3 1330h POSTER

An $^{40}Ar/^{39}Ar$ Plateau age and Geochr
Proterozoic Intrusives of the Purcell S
Basin, Canada

A M Ghazi (Dept. Geol., Georgia St. Un
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R J Goble and S B Treves (Dept. of Geol.
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M J Wampler (Earth and Atmospheric Sci.,
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A 1400 Ma $^{40}Ar/^{39}Ar$ plateau age for a biotit
a sill is interpreted as the minimum age of en
porphyritic intrusion. They occur as dikes an
upper and lower 50-60 meters of the Appekun
Formations, respectively, and contain megacry
plagioclase (An_{48-52}), Ti-augite, biotite and min

Diabasic intrusions are subdivided into type "A
comprising an extensive suite of mafic rocks
or more sills in Siyeh Formation
separate ages of 70
localities

MANTLE-DERIVED PARENT MAGMAS FOR MASSIF ANORTHOSITES??? J. Longhi (Lamont-Doherty Earth Observatory, Palisades, NY 10964, USA) and J. Vander Auwera (LA Géologie, Pétrologie, Géochimie - B20 - Université de Liège, B-4000, Sart Tilman, Belgium)

The Sr, Nd, and Pb isotopic compositions of massif anorthosites suggest an ultimate mantle origin with a variable signature of crustal contamination [e.g.1]. Recent experimental work [2] has shown that the average high-Al gabbro composition at Harp Lake [3] is multi-saturated (opx+cpx+plag on the liquidus) in the range of 11-12 kbar with opx compositions similar to those of the opx megacrysts and plagioclase similar to the bulk of the plagioclase in the complex. More recent work on a jotunite (Tj) from the Rogaland Province has shown that this rock is close to being a suitable parent for the Bjerkreim-Sokndal (BkSk) layered intrusion [4] — the ideal parent would slightly more magnesian and calcic. Data from this work also show that Tj is multi-saturated with opx+cpx+plag at ~ 13 kbar, and that opx (En_{67}, 600 ppm Cr) and plag (An_{42}) compositions at this pressure are in the range of the more evolved opx megacryst and plagioclase "phenocryst" compositions observed within the massif anorthosites [5]. Thus liquids similar in composition to HLCA and a slightly more primitive Tj may have been residual to the early stages of anorthosite petrogenesis in the lower crust. A constraint for mantle derivation is that anorthosites' parent magma compositions lie along liquid lines descent for primitive basalts. We have shown previously that this is not the case for Tj: Tj is nearly coplanar with its liquidus plagioclase and pyroxene compositions at 13 kbar, implying that Tj lies in or near a thermal divide at lower crustal pressures [4]. Re-examination of the data of [2] shows a similar situation for HLCA at 11.5 kbar. Thus neither composition is likely to be reached by fractionation or fractionation plus assimilation (AFC), because assimilation requires crystallization to provide the heat of melting which will drive the liquid composition away from the thermal divide. At lower pressures the pyx+plag thermal divide disappears, so AFC processes can produce compositions similar to HLCA and Tj, but the range of primitive basalts is highly restricted because another thermal divide, ol+aug+plag, develops as soon as pyx+plag disappears. The coincidence of AFC yielding significant amounts of liquid that both lies in a thermal divide at a higher pressure and is appropriately multi-saturated at the same higher pressure is rendered even less probable by the requirement for AFC to take place in the cooler middle crust.

It thus appears that the lower crustal fusion hypothesis of [6] is most consistent with the data: Large ponded intrusions of basalt provide the heat for an initial stage of melting which extracts a rapakivi granite component from recently accreted [1] lower crust and leaves an anhydrous plagioclase + pyroxene residuum enriched in Sr and Eu relative to Rb and the REE; continued heating produces a second stage of melt that lies close to or within the pyroxene+plagioclase thermal divide and that has at best a small negative, or perhaps even positive, Eu-anomaly. Modest degrees of partial melting produce liquids with the major element composition of high-Al basalt — such liquids are parental to anorthosites with more calcic plagioclase (e.g., Harp Lake [3]); lower degrees of melting produce liquids with higher concentrations of K, Ti, P, and other incompatible elements — such liquids (primitive jotunites) are parental to anorthosites with more sodic plagioclase (e.g., portions of the Rogaland massifs [5]). However, there remains a need for mechanical enrichment of the parental liquids with plagioclase crystals in order to produce anorthositic rocks, because cotectic melts of the lower crust will have Al_2O_3 concentrations much lower than anorthosites.

[1] Demaiffe D., Weiss D., Michot J., and Duchesne J.C. (1986) *Chem. Geol. 57,* 167-179. [2] Fram M.S. and Longhi J. (1992) *Am. Mineral. 77,* 605-616. [3] Emslie, R.F. (1980) *Geol. Surv. Can. Bull. 293,* 1-136. [4] Vander Auwera J. and Longhi J. (1994) *Contrib. Mineral. Petrol.*, in press. [5] Duchesne J.C. and Maquil R. (1992) in *The Rogaland Intrusive Massifs,* J.C. Duchesne (ed.), IGCP-290, pp. 7-17. [6] Taylor S. R. and McClennon S. M. (1985) *The Continental Crust:*. Blackwell.

DIFFERENTIATES OF THE PICRITIC GLASS MAGMAS: THE MISSING MARE BASALTS;

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One prediction of the hypothesis that mare basalts are differentiates of more primitive magmas similar in composition to the picritic volcanic glasses (1,2,3) is that there should be mare basalts with compositions spanning the range expected of differentiates of the known picritic compositions. Figure 1 demonstrates that this predicition has not yet been confirmed. Figure 1 shows the TiO_2 and MgO concentrations of fine-grained mare basalts and picritic glasses (2); also plotted are calculated differentiation trends of a few of the picritc glass compositions. It is clear that there are no basalt compositions near the kinked portions of the red and black glass trends. There is also a dearth of basalt compositions corresponding to differentiates of the yellow glasses. Basalts with intermediate and very high TiO_2 concentrations may thus be thought of as "missing", at least from the sample collections. Remote sensing studies (4,5) indicate abundant unsampled flows with intermediate TiO_2, so the non-primary basalt hypothesis is at least consistent with the data in this portion of the composition range and the dearth of the mare basalts with intermediate TiO_2 is apparently a sampling problem. The missing very-high-Ti basalts are more problematical. The same remote sensing studies also describe extensive areas of an unsampled high-Ti basalt type (HDSA) that is similar to Apollo 11 and 17 basalts in TiO_2 concentration,but is apparently more mafic. Taken at face value, the reflectance spectra of these basalts (actually soils) do not correspond the differentiates of the red and black glasses, which should have TiO_2 concentrations in the range of 11(evolved) to 17 (primitive) wt%. So it is not clear whether the very-high-Ti basalts are simply missing from our collections or are simply not present on the Moon.

Given the possibility that a few rare fragments of these intermediate-Ti and very-high-Ti basalts may be present in our collections, it should be useful to describe some the petrographic and chemical characteristics predicted for differentiates of yellow, red, and black glass compositions that might help in their discovery. Accordingly, I have calculated near-perfect fractional crystallization sequences for the Apollo 14 and 15 Yellow, the Apollo 15 Red, and the Apollo 14 Black glass compositions tabulated by (2). Partial results are listed in Table 1 in terms of temperature, volume (approximate) percent crystallized, order of mineral appearance, and silicate mineral composition. The calculations were arbitrarily stopped at 80 mole% crystallization.

Intermediate-Ti Basalts: Calculations show that all picritic glass compositions crystallize olivine and chromite early. The result in fine-grained differentiates should be olivine phenocrysts with chromite inclusions. After olivine and chromite, differentiates of the Apollo 14 and 17 yellow glass magmas should differ only slightly in crystallization sequence. Augite will crystallize next in the Apollo14 magma followed soon after by nearly simultaneous crystallization of plagioclase and ilmenite. The result in fine- to intermediate-grained rocks should be a sub-ophitic texture with ilmenite having the same textural position as plagioclase. Although plagioclase should be the second major phase to crystallize from the Apollo 17 magma (assuming no nucleation delays), plagioclase, augite, and ilmenite should begin to crystallize nearly simultaneously and produce an ophitic groundmass. In general appearance, differentiates of the Apollo 14 and 17 yellow glass magmas will resemble Apollo 12 ilmenite basalts (6), but with 1 to 3 % more ilmenite in the mode.

Very-high-Ti Basalts. Mineralogically, differentiates of the Apollo 15 Red and 14 Black glass magmas should be generally similar to some of the Apollo 11 and 17 basalts: phenocrysts of olivine with chromite inclusions and phenocrysts of ilmenite with cores of armalcolite in groundmasses of intergrown augite, plagioclase, and ilmenite. It is possible that very fine-grained or vitrophyric basalts might have phenocrysts of only olivine (with chromite) and armalcolite. Chemically, these basalts would be readily distinguishable from known mare basalts by their concentrations of TiO_2: up to 15 wt% (Red) or 17 wt% (Black). Petrographically, the total mode of opaques (>25%) would be diagnostic in medium- to coarse-grained rocks, and in fine-grained rocks the amount of armalcolite might approach 10%.

An interesting feature of these calculations is that pigeonite does not appear to crystallize from any of the four magmas modeled. Although pigeonite is the dominant pyroxene in low-Ti mare basalts, it is not a ubiquitous phase in the Apollo 12 ilmenite basalts (6) and in the high-Ti basalts (e.g.,7). Its absence in these latter rocks and in the calculations can be explained in terms of the extent of fractionation which is a reflection of cooling rate. Melting experiments on lunar compositions have shown that pigeonite is unstable with respect to olivine plus silica in highly ferroan ($Mg' < 0.15$) liquids (8). If Mg' in a magma drops below this value before the magma reaches saturation with pigeonite, then no pigeonite will crystallize. This situation is most favorable in magmas undergoing near perfect fractional crystallization that are saturated with augite because augite crystallization will cause Mg' to decrease with minimal increase in SiO_2. The same magma undergoing inefficient fractionation will crystallize pigeonite because Mg' decreases much more slowly per unit of crystallization and the magma is able to reach a sufficiently high SiO_2 content to stablize pigeonite while Mg' is still above 0.15. Thus it is observed that pigeonite is stable predominantly in the more coarsely grained and hence more slowly cooled members of the Apollo 12 ilmenite basalt suite (6) and Apollo 17 high-Ti basalts (7). A related feature of the calculations is the continuous stability of olivine (Table 1). Once augite begins to crystallize the proportion of olivine crystallizing falls almost to zero and then increases once silica begins to crystallize. If pigeonite were to crystallize, then olivine would quickly disappear only to reappear in the latest stages of crystallization when pigeonite became unstable.

REFERENCES

(1) Taylor S.R. (1982) *Planetary Science: A Lunar Perspective*, The Lunar & Planetary Institute. (2) Delano J.W. (1986) *J. Geophys. Res., 91*, D201-D213. (3) Longhi J. (1987) *J. Geophys. Res., 92*, E349-E360. (4) BVSP (1981) *Basaltic Volcanism on the Terrestrial Planets*, The Lunar & Planetary Institute. (5) Pieters C.M., Head J.W., Adams J.B., McCord T.B., Zisk S., and Whitford-Stark J.L. (1980) *J. Geophys. Res., 85*, 3913-3938. (6) Dungan M.A. and Brown R. W. (1977) *Proc. Lunar Sci. Conf. 8th*, p. 1339-1381. (7) Longhi J. (1974) *Proc. Lunar Sci. Conf. 5th*, p. 447-469. (8) Longhi J. and Pan V. (1988) *J. Petrol., 29*, 115-147.

TABLE 1. Fractional Crystallization of Picritic Magmas

Apollo 14 Yellow

T(°C)	VFX	
1350	0	fo_{78}
1346	1	fo_{78}, crmt
1126	39	fo_{54}, (crmt), $Wo_{35}En_{39}$
1091	46	fo_{48}, $Wo_{34}En_{35}$, An_{88}
1084	50	fo_{46}, $Wo_{34}En_{33}$, An_{87}, ilm
1043	83	fo_{16}, $Wo_{35}En_{14}$, An_{80}, ilm

Apollo 15 Red

T(°C)	VFX	
1254	0	fo_{80}
1248	1	fo_{80}, crmt
1207	10	fo_{77}, crmt, arm
1136	30	fo_{67}, (arm), ilm
1132	31	fo_{66}, ilm, $Wo_{33}En_{46}$
1115	37	fo_{63}, ilm, $Wo_{33}En_{44}$,An_{88}
1049	84	fo_{28}, ilm, $Wo_{36}En_{22}$,An_{78}

Apollo 17 Yellow

T(°C)	VFX	
1292	0	fo_{78}
1287	1	fo_{78}, crmt
1126	28	fo_{62}, (crmt), An_{92}
1123	31	fo_{61}, An_{92},$Wo_{34}En_{43}$
1119	37	fo_{59}, An_{91},$Wo_{33}En_{42}$,ilm
1069	76	fo_{32}, An_{86},$Wo_{34}En_{26}$,ilm, sil
1064	84	fo_{21}, An_{83},$Wo_{33}En_{19}$,ilm, sil

Apollo 14 Black

T(°C)	VFX	
1273	0	fo_{80}
1268	1	fo_{80}, crmt
1244	8	fo_{78}, crmt, arm
1172	36	fo_{67}, (arm), ilm
1129	43	fo_{61}, ilm, $Wo_{34}En_{42}$
1092	56	fo_{52}, ilm, $Wo_{35}En_{36}$,An_{90}
1060	78	fo_{33}, ilm, $Wo_{39}En_{25}$,An_{87}, sil
1047	84	fo_{25}, ilm, $Wo_{39}En_{20}$,An_{86}, sil

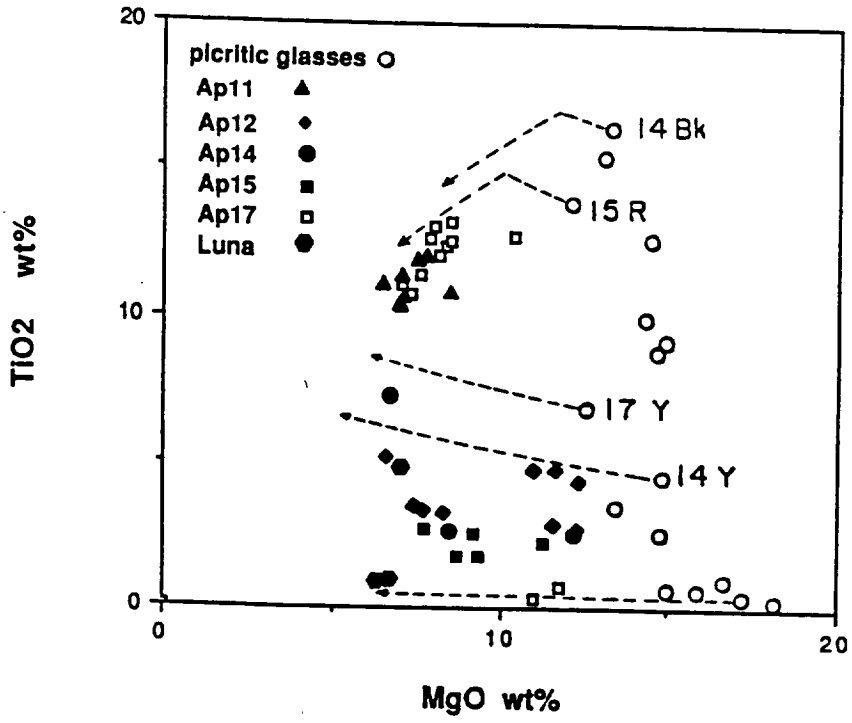


Figure 1. TiO_2 versus MgO in mare basalts and picritic glasses. Dashed lines are partial liquid lines of descent calculated by the method of (3). For green and yellow glasses, liquids are saturated only with olivine and chromite; for red and black glasses, negative slope represents liquids saturated with olivine and chromite, positive slope indicates olivine, chromite, and armalcolite.

COMPLEX MAGMATIC PROCESSES ON MARS; J. Longhi, Lamont-Doherty Geological Observatory, Palisades, NY 10964

The SNC (shergottites-nakhlites-Chassigny) meteorites reveal a surprising diversity of magmatic styles on their parent body, which the weight of evidence suggests is Mars (1). There is evidence for large scale mantle heterogeneities, multi-stage melting, extreme fractionation of REE, assimilation of a long-term light REE-enriched component (a 'granitic' crust?), mantle metasomatism, and possibly CO_2-fluxed melting. In some respects the style of martian magmatism is intermediate between that of the Moon and the Earth, however, the terrestrial end-member has more of the character of hot-spots and continental rifts than mid-ocean ridges.

Estimates of the major element composition of the SNC parent magmas show them to be hpyersthene-normative, high-Fe, low-Al liquids (2). As might be anticipated, calculated densities of these liquids are high (2.75-2.96) and viscosities are low (4-128 poise). Figure 1 illustrates these liquid compositions projected from the Olivine component onto the Opx-Pl-Wo plane. The liquidus boundaries are appropriate for this general class of compositions at low-pressure; arrows show direction of decreasing temperature. The diagram is consistent with the petrography: olivine and pyroxene crystallize early, plagioclase crystallizes late (1); this crystallization pattern is different from terrestrial MORB's and continental tholeiites in which plagioclase crystallizes early. The range of projected SNC parent magma compositions is similar to but somewhat more extensive than those of lunar mare basalts and terrestrial basaltic komatiites; the shergottite members of the suite are also generally similar to the soil composition (V) at the Viking Lander sites (3). The shaded area in Figure 1 represents the range of liquids produced by a few percent partial melting of a four-phase terrestrial mantle peridotite (ol-opx-cpx-Al-phase [plag/sp/gar]) at high pressure. Both increasing pressure and increasing alkalis shift liquids to the right in this field. A martian mantle melt (BH) in equilibrium with ol-opx-cpx-gar at 23 kb (4) also lies within this field. In order to produce liquids that project into the area of the SNC liquids, it is necessary to do large degrees of partial melting (~50%) of relatively undepleted mantle or smaller degrees of partial melting of a source depleted in Al. Geochemical and isotopic data discussed below show the latter to be the case. This depletion may be accomplished by the accumulation of olivine and pyroxene (lunar style) or extraction of basalt (terrestrial style).

Figure 2 illustrates some important aspects of SNC trace element and isotopic composition. Figure 2A shows the REE concentrations of two calculated parent liquid compositions for Nakhla. Chassigny parent liquids have similar patterns. Nakhla is an unusual rock consisting of large cumulus augite and minor olivine crystals set in a rapidly crystallized matrix (5). The 'closed system' calculation assumes that no net changes in the intercumulus liquid took place after accumulation of the pyroxene, but does allow for partial equilibration of the cumulus crystals and trapped liquid. The 'open system' calculation is a direct calculation based upon the composition of Nakhla augite (6) and the partition coefficients of (7); this calculation allows for the possibility of migration of intercumulus liquid. Both patterns are similar and show dramatic light REE enrichment. This fractionation is truly remarkable in light of the ε_{Nd} value of +16 (6) which requires that the source had a long term pattern of light REE depletion, i.e. something similar to the EETA79001A pattern in Fig. 2B. Compounding the situation is the low Al content of the Nakhla parent liquid (N) evident in Fig. 1 (Pl is the Al bearing component). The Al content is sufficiently low that garnet, which is the most effective REE fractionating agent, cannot have been a residual phase in the parent magma's source region; niether is there much allowance for removal of augite at low pressure. The problem of deriving strongly light-REE enriched magmas from light-REE depeleted source regions is common to terrestrial hot spots, such as Hawaii (8). Single-stage models require prohibitively small degrees of partial melting (<1%), so multi-stage melting models have been invoked to spread the REE fractionation over two or more steps (e.g., 9). Some sort of multi-stage melting process thus seems necessary to explain the Nakhla parent magma composition with the condition that garnet not have been a residual phase in the last stage of melting. The unusually high Wo content of N (Fig. 1) requires either that the source was dominated by augite or that CO_2, which has the potential of drastically increasing the CaO content of melts coexisiting with olivine and pyroxene, fluxed the melting at pressures > 25 kb (2). Since partial melting of pyroxene dominated sources produces small negative Eu anomalies in the liquid (10) and since there is no evidence of such an anomaly in Fig. 2A, the presence of CO_2 in martian melting processes must be seriously considered.

Fig. 2B illustrates very different REE patterns for the shergottites. The Shergotty pattern is the 70% ICM model taken from (11). The EETA79001A pattern is the bulk rock analysis of (12). EETA79001A is a fine-grained basaltic rock with 10-15 % mafic xenocrysts (13). These xenocrysts will likely have only a minor diluting effect on incompatible elements, so the pattern in Fig. 2B is believed to close to, albeit slightly lower and steeper than, the true parent liquid pattern. The crystallization ages of the shergottites are controversial because of variable shock effects on the isotopic systems and consequentlythe values of ε_{Nd} are model dependent. The values shown in Fig. 2B are consistent with the 180 m.y. age advocated by (14). This age is chosen here because only the younger ages, which yield $\varepsilon_{Nd} > 0$, are petrologically reasonable, and because the 350 m.y. age reported by (15) has been shown to be a mixing line (11). Given these qualifications, projection point (E_g in Fig. 1), the REE pattern, and ε_{Nd} of EETA79001A have a straightfoward explanation: partial melting of a low-Al source region with a long term light-REE depletion. In this regard, the source region was similar to that of Nakhla although the EETA79001A magma genesis was apparently much simpler. The ε_{Nd} values for Nakhla and EETA79001A are much higher than typical terrestrial basaltic values, but are more typical of lunar mare basalts. This similarity suggests that Mars was more

12A.

like the Moon in its ability to maintain long term isotopic heterogeneities in its mantle. Lack of crustal recycling on Mars and/or less vigorous mantle convection than the Earth are probable explanations.

Given the similarity of mineral compositions in Shergotty to those in the groundmass of EETA79001A, it is likely that their parent magmas lay along similar liquid lines of descent, as suggested by Fig. 1, and hence they were derived from similar primary magmas and source regions. If so, then REE pattern and ε_{Nd} of the Shergotty parent magma in Fig. 2B are readily explicable as those of a magma derived from a depleted source region like EETA79001A, but subsequently contaminated by a low-temperature, long-term, light-REE enriched component. The slight U-shape in the light REE is especially indicative of such a contamination. This component probably is crustal, but whether it is older basalt, like the Nakhla parent magma (Fig. 2A), or 'granitic' is not clear; the physics of assimilation favors an evolved composition with a low melting point, however. One thing that is clear is the absence of a negative Eu-anomaly in the Shergotty REE pattern. Consequently, this crustal component was unlike lunar KREEP, which has a prominent negative Eu-anomaly (16).

REFERENCES: (1) McSween, H. Y. (1985) *Revs. Geophys., 23*, 391-416. (2) Longhi, J. and Pan, V. (1989) *Proc. Lunar Planet. Sci. Conf. 19th*, p. 451-464. (3)Toulmin, P., III, Baird, A. K., Clark, B. C., Keil, K., Rose, H. J., Jr., Christian, R. P., Evans, P. H., and Kelliker, W. C. (1977) *J. Geophys. Res. 82*, 4625-4634. (4) Bertka, C. M. and Holloway, J. R. (1988) *Proc. Lunar Planet. Sci. Conf. 18th*, 723-739. (5) Treiman A. H. (1986) *Geochim. Cosmochim. Acta, 50*, 1061-1070. (6) Nakamura, N., Unruh, D. M., Tatsumoto, M. and Hutchinson, R. (1982) *Geochim. Cosmochim. Acta, 46*, 1555-1573. (7) McKay G., Wagstaff J., and Yang S.-R. (1986) *Geochim. Cosmochim. Acta, 50*, 927-937. (8) Chen C.-Y. and Frey F.A. (1985) *J. Geophys. Res., 90*, 8743-8768. (9) Ribe N.M. (1988) *Earth Planet. Sci. Lett., 88*, 37-46. (10) Shearer C.K. and Papike J.J. Proc. (1989) *Lunar and Planet. Sci. Conf. 20th*, in press. (11) Lundberg L.L, Crozaz G., McKay G., and Zinner E. (1988) *Geochim. Cosmochim. Acta, 52*, 2147-2163. (12) Burghele A., Dreibus G., Palme H., Rammensee W., Spettel B., Weckwerth G., and Wanke H. (1983) *Lunar Planetary Science XIV*, 80-81. (13) McSween H.Y. and Jarosewich E. (1983) *Geochim. Cosmochim. Acta, 47*, 1501-1513. (14) Jones J.H. (1986) *Geochim. Cosmochim. Acta, 50*, 969-977. (15) Jagoutz, E. and Wänke, H. (1986) *Geochim. Cosmochim. Acta, 50*, 939-953. (16) Warren P. H. and Wasson J.T. (1979) *Rev. Geophys. Space Phys., 17*, 73-88.

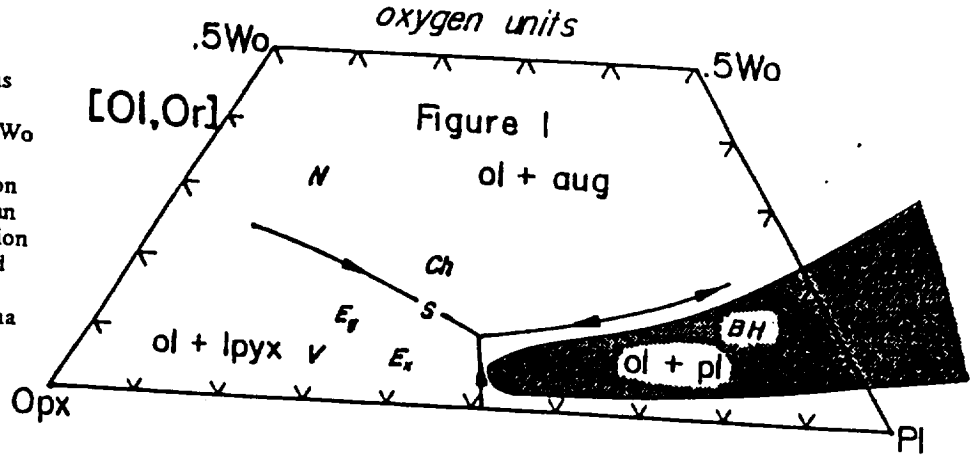
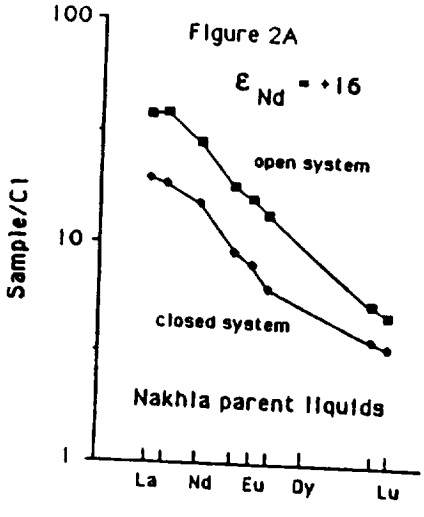
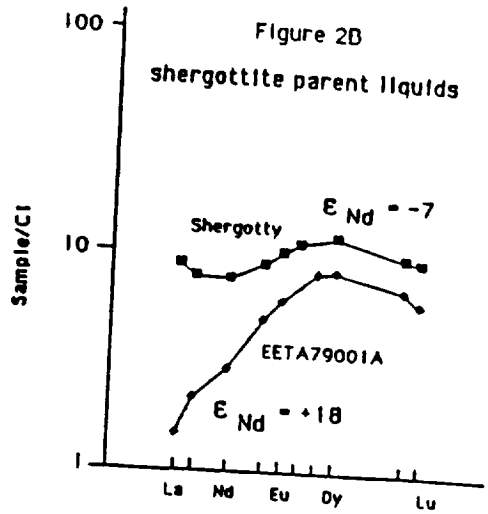


Figure 1. Compositions of SNC parent magmas projected onto Opx-Pl-Wo plane after (2). V is Viking soil composition (3); BH is 23 kb martian primary melt composition (4). Figure 2. Calculated REE in Nakhla and shergottite parent magma compositions.





MAGMATIC PROCESSES ON MARS: INSIGHTS FROM SNC METEORITES;

J. Longhi, Lamont-Doherty Geological Observatory, Palisades, NY 10964.

The composition and petrology of the SNC (shergottites-nakhlites-Chassigny) meteorites reveal a surprising diversity of magmatic processes on their parent body, which the weight of evidence suggests is Mars (1). There is evidence for large scale mantle heterogeneities, multi-stage melting, extreme fractionation of REE, assimilation of a long-term light REE-enriched component (a 'granitic' crust?), mantle metasomatism, and possibly CO_2-fluxed melting. In some respects the style of martian magmatism is intermediate between that of the Moon and the Earth, with the terrestrial component having some of the geochemical character of hot-spot and arc basalts.

Estimates of the major element composition of the SNC parent magmas show them to be hpyersthene-normative, high-Fe, low-Al liquids (2). As might be anticipated, calculated densities of these liquids are high (2.75-2.96) and viscosities are low (4-128 poise). These compositions are consistent with the petrography: olivine and pyroxene crystallize early, plagioclase crystallizes late (1); this crystallization pattern is different from terrestrial MORB's and continental tholeiites in which plagioclase crystallizes early. These low-Al liquids require either large degrees (~50%) of partial melting of an undepleted source or smaller degrees of melting of a depleted source. Geochemical and isotopic data discussed below show the latter certainly to be the case for Nakhla and probably to be the case for the shergottites. This depletion may be accomplished by the accumulation of olivine and pyroxene (lunar style) or extraction of basalt (terrestrial style).

Figure 1 illustrates some important aspects of SNC trace element and isotopic composition. Figure 1A shows the REE concentrations of two calculated parent liquid compositions for Nakhla. Chassigny parent liquids have similar patterns. Nakhla is an unusual rock consisting of large cumulus augite and minor olivine crystals set in a rapidly crystallized matrix (5). The 'closed system' calculation assumes that no net changes in the intercumulus liquid took place after accumulation of the pyroxene, but does allow for partial equilibration of the cumulus crystals and trapped liquid. The 'open system' calculation is a direct calculation based upon the composition of Nakhla augite (6) and the partition coefficients of (7); this calculation allows for the possibility of migration of intercumulus liquid. Both patterns are similar and show dramatic light REE enrichment. This fractionation is truly remarkable in light of the ε_{Nd} value of +16 (6) which requires that the source had a long term pattern of light REE depletion, i.e. something similar to the EETA79001A pattern in Fig. 2B. Compounding the situation is the low Al content of the Nakhla parent liquid (N) evident in Fig. 1 (Pl is the Al bearing component). The Al content is sufficiently low that garnet, which is the most effective REE fractionating agent, cannot have been a residual phase in the parent magma's source region; niether is there much allowance for removal of augite at low pressure. The problem of deriving strongly light-REE enriched magmas from light-REE depeleted source regions is common to terrestrial hot spots, such as Hawaii (8). Single-stage models require prohibitively small degrees of partial melting (<1%), so multi-stage melting models have been invoked to spread the REE fractionation over two or more steps (e.g., 9). Some sort of multi-stage melting process thus seems necessary to explain the Nakhla parent magma composition with the condition that garnet not have been a residual phase in the last stage of melting. In addition to low-Al the Nakhla parent magma also had an unusually high concentration of CaO (~14 wt% (2)). The combination of low-Al and high-Ca requires either that the source was dominated by augite or that CO_2, which has the potential of drastically increasing the CaO content of melts coexisiting with olivine and pyroxene, fluxed the melting at pressures > 25 kb (2). Since partial melting of pyroxene dominated sources produces small negative Eu anomalies in the liquid (10) and since there is no evidence of such an anomaly in Fig. 2A, the presence of CO_2 in martian melting processes must be seriously considered.CO_2 is also an effective carrier of light REE (11), so CO_2 may have affected both trace and major elements during melting.

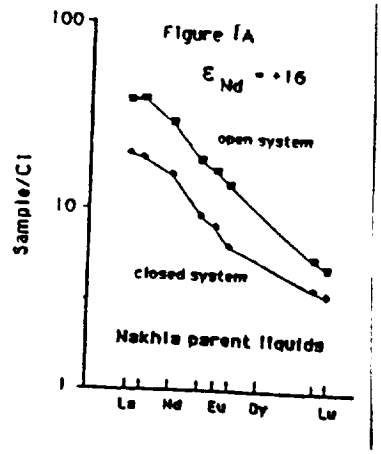
Fig. 1B illustrates very different REE patterns for the shergottites. The Shergotty pattern is the 70% ICM model taken from (12). The EETA79001A pattern is the bulk rock analysis of (13). EETA79001A is a fine-grained basaltic rock with 10-15 % mafic xenocrysts (14). These xenocrysts will likely have only a minor diluting effect on incompatible elements, so the pattern in Fig. 1B is believed to close to, albeit slightly lower and steeper than, the true parent liquid pattern. The crystallization ages of the shergottites are controversial because of variable shock effects on the isotopic systems and consequentlythe values of ε_{Nd} are model dependent. The values shown in Fig. 1B are consistent with the 180 m.y. age advocated by (15). This age is chosen here because only the younger ages, which yield $\varepsilon_{Nd} > 0$, are petrologically reasonable, and because the 350 m.y. age reported by (16) has been shown to be a mixing line (12). Given these qualifications, the low-Al content, the depleted light-REE pattern, and $+\varepsilon_{Nd}$ of EETA79001A have a straightfoward explanation: partial melting of a low-Al source region with a long term light-REE depletion. In this regard, the source region was similar to that of Nakhla although the EETA79001A magma genesis was apparently much simpler. The ε_{Nd} values for Nakhla and EETA79001A are much higher than typical terrestrial basaltic values, but are more typical of lunar mare basalts. This similarity suggests that Mars was more like the Moon in its ability to maintain long term isotopic heterogeneities in its mantle. Lack of crustal recycling on Mars and/or less vigorous mantle convection than the Earth are probable explanations.

Given the similarity of mineral compositions in Shergotty to those in the groundmass of EETA79001A, it is likely that their parent magmas lay along similar liquid lines of descent, as suggested by Fig. 1, and hence they were derived from similar primary magmas and source regions. If so, then REE pattern and ε_{Nd} of the Shergotty

Shergotty parent magma in Fig. 2B are readily explicable as those of a magma derived from a depleted source region like EETA79001A, but subsequently contaminated by a low-temperature, long-term, light-REE enriched component. The slight U-shape in the light REE is especially indicative of such a contamination. This component probably is crustal, but whether it is older basalt, like the Nakhla parent magma (Fig. 2A), or 'granitic' is not clear; the physics of assimilation favors an evolved composition with a low melting point, however. One thing that is clear is the absence of a negative Eu-anomaly in the Shergotty REE pattern. Consequently, this crustal component was unlike lunar KREEP, which has a prominent negative Eu-anomaly (17).

Fig. 2A contains REE concentrations for the Nakhla parent ('closed system') and the bulk data for EETA79001A taken from Fig. 1 plus additional calculated and measured concentrations of some high-field-strength elements (HFSE) arranged in order of incompatibility. Fig. 2A shows that there are complementary anomalies for Ta, Hf, and Zr in the Nakhla and EETA79001A patterns, thus supporting the hypothesis that the shergottites were generated by remelting a source depleted in a Nakhla-like component. Fig. 2B schematically illustrates typical incompatible element patterns for basalts from terrestrial oceanic islands (OIB), volcanic arcs (ARC), and mid-ocean ridges (MORB). Despite the fact that there is no evidence of plate tectonics on Mars, the Nakhla parent magma pattern appears more similar to the ARC pattern than to OIB or MORB. This similarity is probably due to similar fractionations of the REE from the HFSE during transport by a CO_2-rich vapor phase, rather than similar tectonic styles.

REFERENCES: (1) McSween, H. Y. (1985) *Revs. Geophys., 23*, 391-416. (2) Longhi, J. and Pan, V. (1989) *Proc. Lunar Planet. Sci. Conf. 19th*, p. 451-464. (3)Toulmin, P., III, Baird, A. K., Clark, B. C., Keil, K., Rose, H. J., Jr., Christian, R. P., Evans, P. H., and Kelliker, W. C. (1977) *J. Geophys. Res. 82*, 4625-4634. (4) Bertka, C. M. and Holloway, J. R. (1988) *Proc. Lunar Planet. Sci. Conf. 18th*, 723-739. (5) Treiman A. H. (1986) *Geochim. Cosmochim. Acta, 50*, 1061-1070. (6) Nakamura, N., Unruh, D. M., Tatsumoto, M. and Hutchinson, R. (1982) *Geochim. Cosmochim. Acta, 46*, 1555-1573. (7) McKay G., Wagstaff J., and Yang S.-R. (1986) *Geochim. Cosmochim. Acta, 50*, 927-937. (8) Chen C.-Y. and Frey F.A. (1985) *J. Geophys. Res., 90*, 8743-8768. (9) Ribe N.M. (1988) *Earth Planet. Sci. Lett., 88*, 37-46. (10) Shearer C.K. and Papike J.J. Proc. (1989) *Lunar and Planet. Sci. Conf. 20th*, in press. (11) Wendlendt R.F. and Harrison W.J. (1979) *Contrib. Mineral. Petrol., 69*, 409-419. (12) Lundberg L.L, Crozaz G., McKay G., and Zinner E. (1988) *Geochim. Cosmochim. Acta, 52*, 2147-2163. (13) Burghele A., Dreibus G., Palme H., Rammensee W., Spettel B., Weckwerth G., and Wanke H. (1983) *Lunar Planetary Science XIV*, 80-81. (14) McSween H.Y. and Jarosewich E. (1983) *Geochim. Cosmochim. Acta, 47*, 1501-1513. (15) Jones J.H. (1986) *Geochim. Cosmochim. Acta, 50*, 969-977. (16) Jagoutz, E. and Wänke, H. (1986) *Geochim. Cosmochim. Acta, 50*, 939-953. (17) Warren P. H. and Wasson J.T. (1979) *Rev. Geophys. Space Phys., 17*, 73-88.



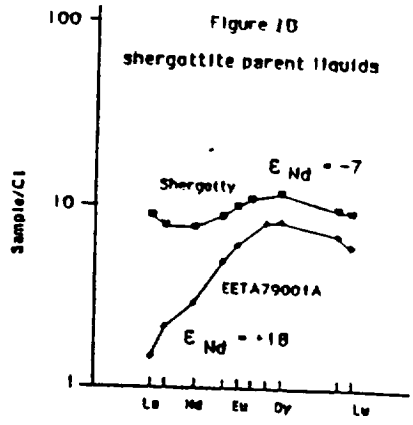
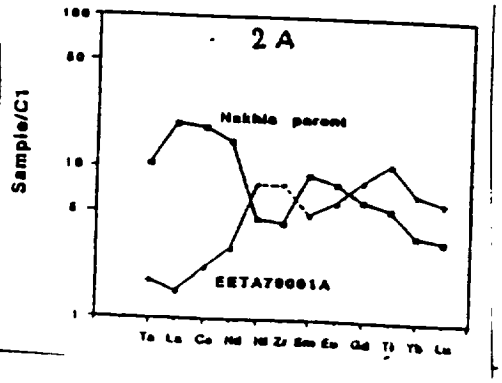


Figure 1. REE in Nakhla and shergottite parent magma compositions. Nakhkla - calc.,this study; Shergotty - calc(12); EETA79001A - bulk (13).



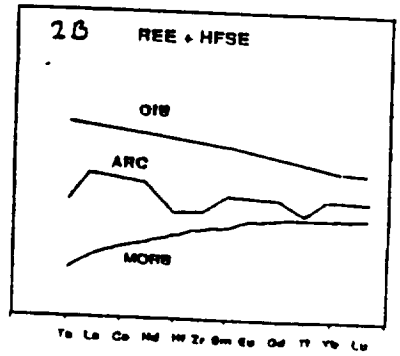


Fig. 2 A. Calculated REE and HFSE in Nakhla parent magma and bulk EETA79001A. B. Typical (schematic) patterns for terrestrial basalts.

EXPERIMENTAL PETROLOGY OF DIKES RELATED TO ANORTHOSITE COMPLEXES;

M.S. Fram and J. Longhi (Lamont-Doherty Geological Observatory of Columbia University, Palisades, NY 10964)

One approach to determining the origin of lunar anorthosites is to examine terrestrial analogues and then adapt resulting models to the lunar case. Lunar anorthosites are enigmatic for several reasons: petrologic and geochemical evidence indicate that none of the mafic rocks yet sampled on the Moon were formed in the same episode as the anorthosites (1,2); they are almost pure plagioclase rocks; and associated mafic minerals have fairly low magnesium numbers (2). These features imply extensive fractionation, efficient separation of plagioclase, and a spatial separation from any mafic cumulates produced.

Proterozoic massif anorthosites share many of these features, although there are important differences between the lunar and terrestrial cases. Some massif anorthosite complexes include dikes which may hold clues to the parental magma and thus to the origin of the anorthosites. We have examined the phase equilibria at 1atm and 7-30kb of two of the dike compositions (table 1): an anorthositic dike from the Nain Complex, sample 500B which has a bulk composition similar to the anorthosite plutons (3); and a synthetic high aluminum basalt, sample HLC-A, which resembles the dikes and chilled margins found with the Harp Lake anorthosites (4). Since massif anorthosite complexes contain no hydrous minerals and have anhydrous aureoles, we conducted all experiments under anhydrous conditions. The very high liquidus temperatures (1365°-1420°; 1atm to 20kb) and the poor correspondence between experimental phases and compositions of minerals in the natural rock indicate that 500B cannot be a liquid or a liquid with suspended plagioclase. The 500B experiments form augite as the second phase, whereas the major mafic phase in the natural rock is inverted pigeonite. Furthermore, plagioclase formed in the experiments is more anorthitic (An58-63) than cores in the natural rock (An50-55).

Experiments on the high aluminum basalt sample, HLC-A, are summarized in figure 1. Plagioclase is the liquidus phase from 1atm through 10kb, followed by olivine at low pressures and by orthopyroxene from 7 to 10kb. At 15kb, the liquidus phase is an aluminous pyroxene followed by plagioclase. These results imply that HLC-A is a multiply saturated liquid at ~12kb (figure 1). The compositions of the phases produced in the experiments match those in the Harp Lake anorthosites (figure 2) showing that fractionation of a basaltic liquid can produce the minerals in the anorthosites. The composition of the aluminous pyroxene resembles the bulk composition of pyroxene megacrysts found in many massif anorthosites. Figure 3 shows the dike compositions projected from wollastonite onto the plagioclase-olivine-silica plane along with calculated 1atm and estimated 3kb phase boundaries. The 12kb plagioclase-pyroxene boundary will pass through the HLC-A point. Both 500B and HLC-A lie well within the plagioclase field at low pressures, showing that both are expected to crystallize lots of plagioclase. Unfortunately, simple crystallization of HLC-A upon decompression from 12kb to lower pressure does not produce a sufficient volume of plagioclase (only 10% plagioclase crystallizes before the plagioclase-orthopyroxene cotectic is reached). Thus, in addition to decompression of high pressure cotectic basaltic liquids, accumulation of plagioclase is required

Longhi and Ashwal (5) proposed a model in which mats of plagioclase form in the upper parts of large layered intrusions, detach, and rise diapirically to form anorthosite plutons, leaving mafic cumulates at great depths. Although a poly-baric fractionation model for the production of massif anorthosites on Earth works rather well, the differences between the lunar and terrestrial case must be considered before simply applying the model to the Moon. Since the Moon has only one-sixth the gravitational field of the Earth, processes depending on decompression require a much larger depth interval to occur. However, experiments on lunar mare basalts and volcanic glasses show that they are multiply saturated at 10-25kb (6), indicating source regions at 100-500km depth in the Moon. Transport from great depths on the Moon apparently occurred for mare basalts, so perhaps the model for the production of massif anorthosites from diapiric rise of plagioclase layers in basaltic layered intrusions might be applied to anorthosites on the Moon.

REFERENCES: (1) Norman, M.D. and Ryder, G. (1980) *Proc. Lunar Planet. Sci. Conf. 11th*, p. 317-331. (2) Longhi, J. (1982) *J.Geophys. Res. 87, Suppl.*, p. A54-A64. (3) Weibe, R.A. (1979) *Am. J. Ssi., 279*, p.394-410. (4) Emslie, R.F. (1980) *Geol. Surv. Canada, Bull. 293*, 136 pp. (5) Longhi, J and Ashwal L.D. (1985) *Proc. Lunar Planet. Sci Conf 15th, J Geophys. Res., 90, Suppl.*, P. C571-C584. (6) Delano, J.W. (1986) *J Geophys. Res., 91*, p. D201-D213.

	Nain Dike (natural) 500B	Harp Lake Dike (synthetic) HLC-A
SiO2	55.40	49.54
TiO2	0.42	1.83
Al2)s	24.01	17.34
Cr2O3	0.00	0.06
FeO	2.79	10.80
MgO	1.22	6.41
MnO	0.06	0.17
CaO	9.85	8.76
Na2O	4.56	2.84
K2O	0.76	0.44
P2O5	0.17	0.17

Table 1. Starting compositions used in experiments.

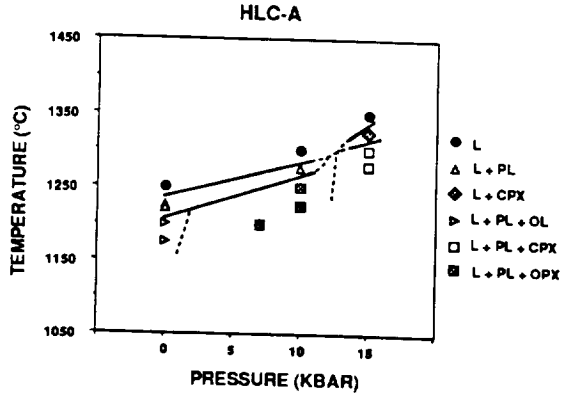


Figure 1. Pressure-Temperature diagram for experiments on HLC-A, the high aluminum basaltic dike composition.

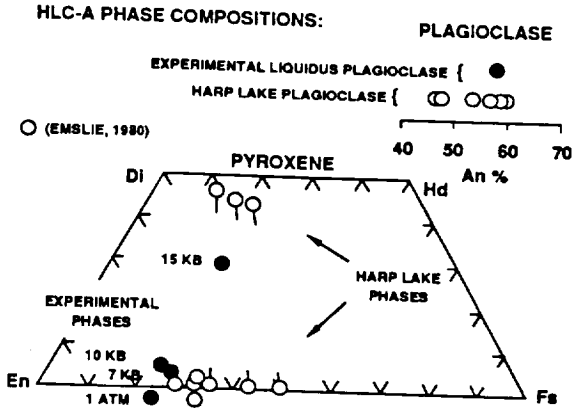


Figure 2. Plagioclase, pyroxene, and olivine compositions from experiments on HLC-A and the Harp Lake anorthosite complex (4).

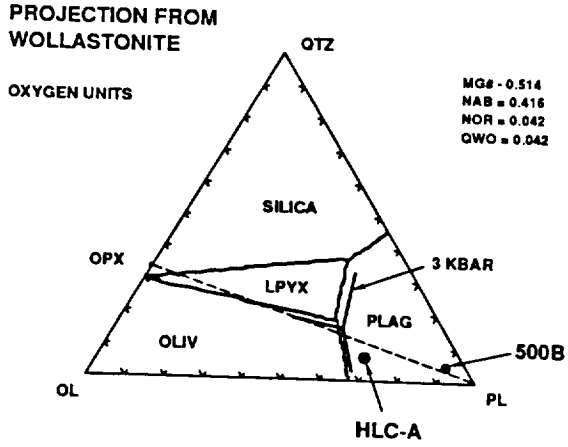


Figure 3. Projection from Wollastonite onto Olivine-Plagioclase-Silica. 1atm phase boundaries calculated for Mg# and normative alkali components of HLC-A. 3kb boundaries are estimated..

DYNAMICAL MELTING MODELS OF MARE BASALTS: J. Longhi, Lamont-Doherty Geological Observatory of Columbia University, Palisades, NY 10964

There has been relatively little attention in recent years given to the processes through which the mare basalts and associated volcanic glasses acquired their major element compositions. Throughout the 1970's petrologists interpreted the point of olivine + pyroxene multiple saturation on the high-pressure liquidus of mare compositions (Fig. 1) as the pressure of melt segregation (e.g. 1,2,3). Although they issued caveats about the assumed primary nature of the compositions they investigated, they did not discuss in any detail the processes of melting or transport to the surface. This latter process has remained something of a mystery because many mare basalts and glasses have implied pressures of segregation > 10 kb corresponding to depths > 200 km. How magma could retain its primary composition through such extended transits to the surface except by kimberlitic eruption from depth has never been clear.

In the 1980's terrestrial geophysicists developed quantitative models for melt migration by porous flow, segregation, and compaction of the crystalline residua in a convecting mantle (4,5,6). These models typically portray a pool of melt consisting of contributions from a range of pressures collected at the top of the ascending mantle column and often ponded at the base of the crust or lithosphere. These treatments have incorporated terms for trace elements with constant partition coefficients, but do not readily lend themselves to major elements whose partition coefficients typically depend on composition. The one attempt to model major elements during polybaric porous flow approximated the process as one in which extraction of melt produced no change in the composition of the continuously melting source (7). This process may be described as total replenishment (TR).

Porous flow models may have some applicability to mare basalt petrogenesis. Mare basalts were produced by partial melting of low-Al mafic to ultramafic sources. During the time of mare petrogenesis the Moon had a thin, but growing elastic lithosphere that eventually supported the loading of mare basalts in the great basins to form mascons: stress modeling suggests elastic lithosphere thicknesses of ~ 50 km at 3.6-3.8 b.y., and increasing to ~ 100 km afterwards (8). Beneath the base of the lithosphere was the mare basalt source region, partially molten in places and probably convecting. In such a physical regime mare basalt magmas would pool at the base of the lithosphere and possibly would rise to the surface along fractures caused by impacts. A major test of the applicability of the porous flow model is its ability to generate composite melts ponded at 100 km or less (< 5 kb) that appear to be multi-saturated at higher pressure and hence to be derived from greater depths. Fig. 2 illustrates this situation: the field of the picritic green volcanic glasses (9) sits astride the 20 kb olivine + orthopyroxene liquidus boundary -- well below the position of the boundary at 5 kb.

To test the porous flow model, I have carried out two variations of polybaric partial melting of lunar compositions and some of the results are shown in Fig. 2. In each case melting begins at some initial pressure and proceeds in a series of steps at progressively decreasing pressure with the melt pooling at some arbitrary lower pressure. One set of calculations involves extraction and accumulation of melt but no change in source composition analogous to (7); these results are labeled TR. The other set involves fractional fusion, i.e. complete extraction of the melt without replenishment; these results are labeled FF. Major element variation during true porous flow, which entails continous interaction between melt and matrix and hence is difficult to calculate, lies somewhere between these tractable extremes. In these calculations I employed the value of 1.2% melt/kb adopted by (7) and based upon the thermal modeling of (10); melting began at pressures indicated in Fig. 2 (25 or 40 kb); melt extraction began at 2.4% (a pressure drop of 2 kb) and continued in 1 or 2 kb steps until 5 kb (or in the case of the 40 kb fractional fusion, until 10kb where Al_2O_3 dropped below 0.2 wt% in the source). In the TR model the amount of melting increases with each 1 kb step (2.4, 3.6, 4.8, etc.); in the FF model the amount of melting remains constant (2.4% per 2 kb step). I derived the model source composition (Table 1) by subtracting 15 wt% Highland Crust from Primitive Mantle (11, Table 8.4). More aluminous compositions shift the FF and TR curves to the right; less aluminous and more ferroan compositions shift these curves to the left.

At first inspection results of these calculations suggest that compositions similar to those of the green glasses may be generated by fractional fusion of an ascending low-Al source that began to melt at great depth (~ 800 km) in the Moon. However, these particular results (FF, 40 kb) do not satisfy the model because pooled melt is too magnesian (Table 1) and the source becomes barren at 200 km. More importantly, the calculations require that the temperature actually increases at pressures less than 30 kb, which is an implausible situation in a convecting mantle. Although this modeling is still at a preliminary stage, it appears that models involving more ferroan source compositions, greater depths for the onset of melting, and a small component of replenishment may satisfy both the compositional and thermal constraints. If such models prove successful, then melting at great depths would correspond more closely to the thermal model of (12) based on 40 ppb of U for the bulk Moon, rather than the model of (13) based on 20 ppb of U, thus implying that the Moon is enriched in refractory elements by a factor of ~ 2 over chondrites.

REFERENCES

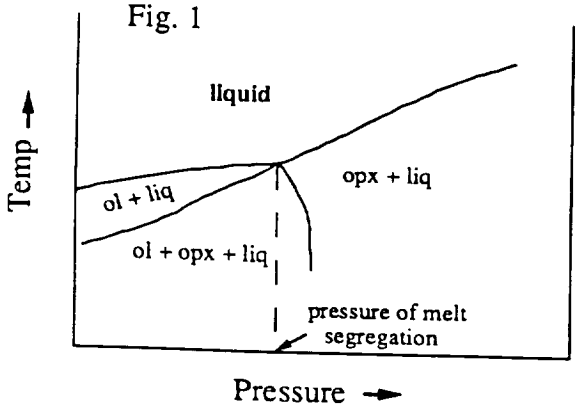
(1) Green D.H., Ware N.G., Hibberson W.O., and Major A. (1971) *Earth Planet. Sci. Lett. 13*, 85-96. (2) Walker D., Longhi J., Stolper E.M., Grove T.L., and Hays J.F. (1976) *Geochim. Cosmochim. Acta 39*, 1219-1235. (3) Delano J.W. (1980) *Proc. Lunar Planet. Sci. Conf. 11th*, p. 251-288. (4) McKenzie D. (1984) J. Petrol. 25, 713-765. (5) McKenzie D. (1985) *Earth Planet. Sci. Lett. 74*, 81-91. (6) McKenzie D. and Richter F. M. (1984) *J. Geol. 92*, 729-740. (7) Klein E.M. and Langmuir C.H. (1987) *J. Geophys. Res. 92*, 8089-8115. (8) Solomon S.C. and Head J.W. (1979) *J. Geophys. Res. 84*, 1667-1682. (9) Delano J.W. (1986) *J. Geophys. Res. 91*, D201-D213. (10) Ahearn J.L. and Turcotte D.L. (1979) *Earth Planet. Sci. Lett. 45*, 115-122. (11) Taylor S.R. Planetary Science: A Lunar Perspective, The Lunar and Planetary Institute, Houston. (12) Solomon S.C. and Longhi J. (1977) *Proc. Lunar Sci. Conf. 8th*, p. 583-599. (12) Hubbard N.J. and Minear J.W. (1975) *Proc. Lunar Sci. Conf. 6th*, p. 1057-1085. (13) Longhi J. and Pan V. (1988) *Proc. Lunar Planet. Sci. Conf. 18th*, p. 459-470.

TABLE 1 (wt%)

	SiO2	TiO2	Al2O3	Cr2O3	FeO	MgO	MnO	CaO	K2O	Na2O	Mg'
model mantle	44.3	0.26	2.88	0.71	11.7	37.3	0.17	2.62	0.002	0.05	0.85
pooled melt 40 kb -FF	45.6	0.77	8.15	0.61	16.6	20.2	0.26	7.74	0.01	0.09	0.68
Ap 15 green glass A (9)	45.5	0.38	7.75	0.56	19.7	17.2	0.22	8.65	0.02	0.16	0.61

FIGURE CAPTIONS

Fig. 1. Schematic high-pressure melting relations of low-Ti mare compositions. Fig. 2. Comparison of green glass compositions (9) and calculated polybaric, pooled melt compositions. Phase boundaries are appropriate for low pressure and compositonal paprameters shown in upper right: Mg' = MgO/(MgO+FeO); NAB and NOR are albite and orthoclase fractions of the normative feldspar; QWo is the Wo coordianate of the Ol-Pl-Wo-Qtz system in mole units (13). FF = fractional fusion model; TR = total replenishment model.



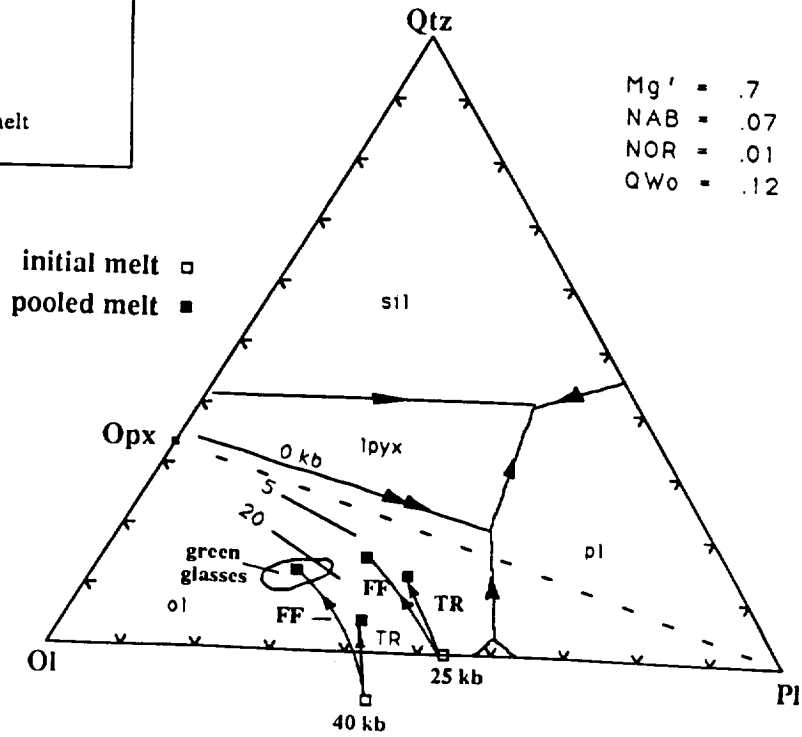


Fig. 2

{Wo}

oxygen units

1115

IMPLICATIONS OF THE VENERA 13 SURFACE ANALYSIS FOR THE CO_2 CONTENT OF THE INTERIOR OF VENUS; J. Longhi[1] and V. Pan[2], 1:Lamont-Doherty Geological Observatory, Palisades, NY 10964; 2: Dept. of Geology, Arizona State University, Tempe, AZ 85287-1404.

Typically, when comparisons of the bulk carbon contents of the Earth and Venus are made, carbon abundances in the Earth's mantle, crust, oceans, atmosphere, and biosphere are integrated, but in the case of Venus only atmospheric carbon (CO_2) is integrated (e.g., 1). These comparisons show the carbon contents of the two planets to be roughly comparable (1.5-4.5 x 10^{-5} kg/kg for the Earth vs 2.7 x 10^{-5} kg/kg for Venus). Explicit assumptions of the estimate for Venus are that little or no carbon remains in the mantle or is bound up in crustal minerals. Given that CO_2 may affect the proportions of major elements in silicate melts at high pressure (2), it is possible to test the first assumption by examining the compositions of Venusian basalts. Unfortunately, we have available only two analyses of basaltic materials. These are surface analyses with large uncertainties obtained from spectrometers on the Venera 13 and 14 landers (3). Because it may be decades before samples are actually returned for high quality analyses, we believe it is worthwhile at least to point out the implications of the data in hand.

The upper mantles of the terrestrial planets are generally believed to consist primarily of a lherzolite assemblage: olivine, orthopyroxene, clinopyroxene ± an aluminous phase (4). The most important effect of CO_2 on high pressure melting is to expand the stability range of orthopyroxene (2). This expansion means that at a given pressure CO_2-bearing melts in equilibrium with olivine, orthopyroxene, and clinopyroxene will have lower SiO_2 and higher CaO contents than similarly saturated CO_2-free melts. In terms of mineral components this means that CO_2-bearing melts saturated with the lherzolite assemblage will have lower proportions of normative orthopyroxene and higher proportions of Ca-silicate components. The effect of H_2O is nearly opposite with H_2O tending to produce SiO_2-rich melts (5). Increasing alkalies at high pressure decrease both the orthopyroxene and Ca-silicate components (6). The effects of CO_2, H_2O, and alkalies are shown as vectors in Fig. 1. Unfortunately, the effects of CO_2 are best known for simpler systems and have not yet been quantified on natural compositions. Nevertheless, Fig. 1 shows a sharp divergence between the "main sequence" of silica-undersaturated terrestrial lavas compiled by (7) and the compositions of experimental liquids in equilbrium with the lherzolite assemblage generated at 10 to 35 kb under fluid-absent conditions (8,9). Given that the release of CO_2-rich volatiles with the eruptions of silica-undersaturated lavas is well documented (10), that carbonates are common accessory minerals in undersaturated rocks (11), and that carbonatites are commonly associated with silica-undersaturated rocks (12), it seems reasonable to attribute the divergence of the main sequence of lava composition from the range of high-pressure fluid-absent melts primarily to the influence of CO_2 .

Fig. 2 contrasts the compositions of the Venera 13 and 14 soils with the main sequence of terrestrial lavas. The Venera 14 analysis is generally similar to terrestrial tholeiites (silica-saturated), however, the Venera 13 analysis, shown as an ellipse to illustrate the uncertianties in SiO_2 and CaO reported by (3) and including a conservative estimate of 1.5 wt % Na_2O, is strongly silica-undersaturated and projects distinctly below the terrestrial main sequence as do the compositions of the fluid-absent high-pressure melts illustrated in Fig. 1. We interpret the Venera 13 composition to imply a potential lack of influence of CO_2 on melting relations in the upper mantle of Venus. If the Venera 13 surface analysis is an accurate (within stated limits) representation of a typical silica-under-saturated volcanic protolith, then the upper mantle of Venus contains much less CO_2 than the Earth's and the assumption that the Venusian mantle is extensively degassed at least with respect to CO_2 seems well-founded.

REFERENCES

(1) Donahue T.M. and Pollack J.B. (1983) in *Venus*, D.M. Hunten, L. Colin, T.M. Donahue, and V.I. Moroz eds., pp. 1003-1036, U. of Arizona Press. (2) Eggler D. (1975) *Geophys. Lab. YB 74*, 468-474. (3) Surkov Yu. A., Moskalyeva, L.P., Shcheglov O.P., Kharyukova V.P., Manvelyan O.S., Kirichenko V.S., and Dudin A.D. (1983) *Proc. Lunar Planet. Sci. Conf. 13*, A481-A493. (4) BVSP (1981) *Basaltic Volcanism on the Terrestrial Planets*, Pergamon. (5) Kushiro I. (1969) *Am. J. Sci. 267A*, 269-294. (6) Pan V. (1989) Ph.D. Thesis, Yale Univ. (7) Pan V. and Longhi J. (1990) *Contrib. Mineral. Petrol. 105*, 569-584. (8) Takahashi E. and Kushiro I. (1983) *Am. Mineral. 68*, 859-879. (9) Falloon T.J. and Green D.H. (1988) *J. Petrol.* Special Lithosphere Issue, 379-414. (11) Gerlach T.M. (1980) *J. Volcanol. and Geothermal Res. 8*, 177-189. (11) Williams H., Turner F.J., and Gilbert C.M. (1982) *Petrography*, 2nd ed., W.H. Freeman and Co. (12) Gittins J. and Tuttle O.F. (1966) *Carbonatites*, Interscience.

Fig. 1

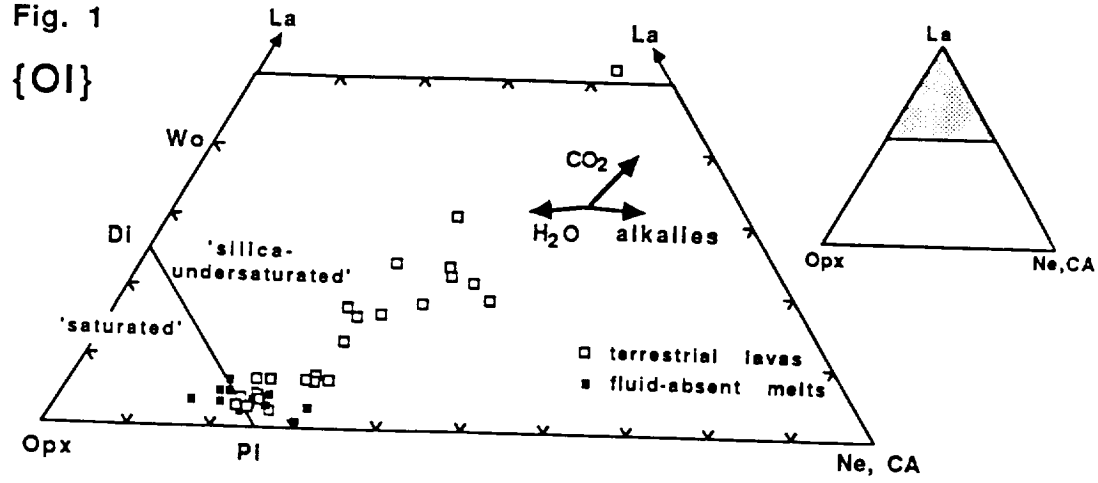


Fig. 2

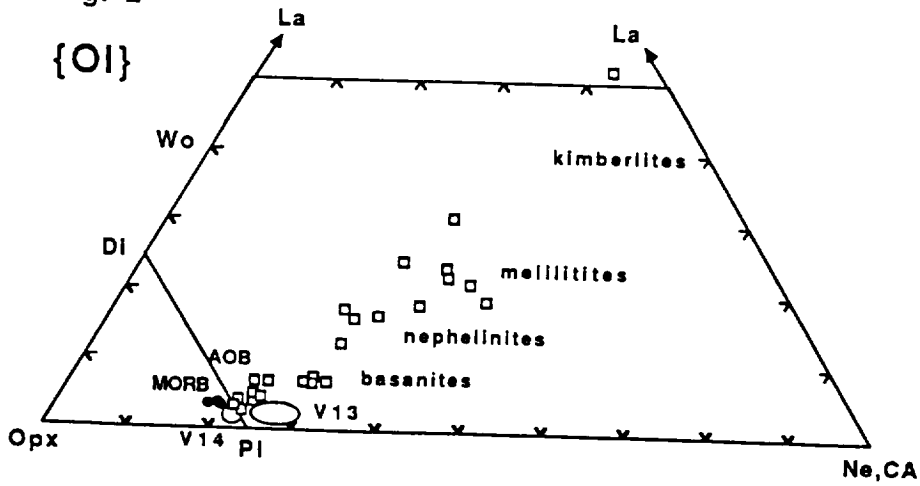


Fig. 1 Comparison of natural terrestrial lavas (7) with high-pressure (10-35 kb) fluid-absent melts in equilibrium with olivine, orthopyroxene, and clinopyroxene (8,9). Vectors show the effects of various components on the projection of compositions of liquids saturated with the lherzolite assemblage. Compositions are projected from the Olivine component (Ol) onto a portion of the plane Orthopyroxene (Opx) - Larnite (La) - Nepheline (Ne), Calcium-aluminate (CA) after (7). The Di-Pl join separates silica-saturated (tholeiitic) from silica-undersaturated compositions. Fig. 2 Comparison of Venera 13 (ellipse) and Venera 14 (circle) surface analyses (3) with terrestrial lava compositions. Projection as in Fig. 1.

THE ORIGIN OF THE GREEN GLASS MAGMAS BY POLYBARIC PARTIAL MELTING: J. Longhi, Lamont-Doherty Geological Observatory of Columbia University, Palisades, NY 10964

One of the important unanswered questions in lunar petrology is how the picritic volcanic glasses (1) acquired and maintained their apparent high-pressure signature. Traditionally, lunar petrologists interpreted the point of olivine + pyroxene multiple saturation on the high-pressure liquidus of mare compositions, illustrated in Fig. 1, as the pressure of melt segregation (e.g., 2,3,4). In the case of the picritic glasses multiple saturation occurs between 17.5 and 25 kb (e.g., 4), implying depths of segregation of 350 to more than 500 km on the Moon. The major alternative to the multiple saturation hypothesis has been invoking relatively large degrees of partial melting at some low pressure that leaves behind a residuum consisting of only olivine. For a source with 2.0 wt% Al_2O_3 and 80% olivine, such a process would require approximately 30% melting to produce a green glass composition; sources with more Al_2O_3 or less olivine would require larger degrees of melting. The prospect of large degrees of melting of sources with small amounts of pyroxene has remained unpopular with petrologists even though they have not explained how a magma could retain its primary composition through transits of several hundred kilometers to the surface: kimberlitic eruptions from depth seem unlikely on the Moon.

In the 1980's terrestrial geophysicists developed quantitative models for melt migration by porous flow, segregation, and compaction of the crystalline residua in a convecting mantle (5,6,7). These models typically portray a pool of melt consisting of contributions from a range of pressures collected at the top of the ascending mantle column and often ponded at the base of the crust or lithosphere. These treatments have incorporated terms for trace elements with constant partition coefficients, but do not readily lend themselves to major elements whose partition coefficients typically depend on composition. Attempts to model major elements during polybaric porous flow approximated the process as one in which extraction of melt produced no change in the composition of the continuously melting source (8,9). However, given that some net separation of melt takes place, some net change in source composition must also take place, so these models should be regarded as endmembers.

Porous flow models may have some applicability to mare basalt petrogenesis. Mare basalts were produced by partial melting of low-Al mafic to ultramafic sources. During the time of mare petrogenesis the Moon had a thin, but growing elastic lithosphere that eventually supported the loading of mare basalts in the great basins to form mascons: stress modeling suggests elastic lithosphere thicknesses of ~ 50 km at 3.6-3.8 b.y., and increasing to ~ 100 km afterwards (10). A thermal lithosphere, probably of comparable thickness, extended the elastic lithosphere. Beneath the base of the lithosphere was the mare basalt source region, partially molten in places and probably convecting. In such a physical regime mare basalt magmas would pool at the base of the lithosphere and possibly would rise to the surface along fractures caused by impacts. A major test of the applicability of the porous flow model is its ability to generate composite melts ponded at ~100 km (5 kb) that appear to be multi-saturated at higher pressure and hence to be derived from greater depths. Fig. 2 illustrates this situation: the field of the picritic green volcanic glasses (1) sits astride the calculated 20 kb olivine + orthopyroxene liquidus boundary -- well below the position of the boundary at 5 kb.

To test the porous flow model, I have calculated numerous variations of polybaric partial melting of lunar mantle compositions designed to simulate porous flow and segregation, using parameterizations of liquidus boundaries based on the high-pressure data of (11). Some of the results are shown in Fig. 2. In each case melting begins at some initial pressure and proceeds in a series of steps at progressively decreasing pressure with the extracted melt pooling in a resevoir; the curved arrows in Fig. 2 show the chemical evolution of the melt pool. The source also becomes progressively depleted in response to melt extraction, but this change is not shown. In these calculations I employed the value of 1.2% melt/kb adopted by (8) and based upon the thermal modeling of (12); melting began at pressures indicated in Fig. 2 (25 or 40 kb); melt extraction began at 2.4% (a pressure drop of 2 kb) and continued in 2 kb steps until 4 kb. To simulate the porous flow process, some pecentage of the melt generated in each step was retained in the source. To achieve a satisfactory fit of ponded melt to green glass composition, I derived the source composition (Table 1) by subtracting 15 wt% Highland Crust and 3% anorthite from Primitive Mantle (13, Table 8.4) and reducing Mg' from 0.85 to 0.80. More aluminous compositions shift the melt resevoir curves to the right; less aluminous and more ferroan compositions shift these curves to the left.

At first inspection results of these calculations (Table 1, Fig. 2) suggest that compositions similar to those of the green glasses may be generated by inefficent fractional fusion of an ascending low-Al source that began to melt at great depth (~ 1000 km) in the Moon. However, the physical aspects of the model need to be explored further. In particular, the model seems to require a discrete melt resevoir that ascends with the melt region. Whether this resevoir can be justified as part of a partially molten diapir or as a porosity wave (14) needs further investigation. Also, the temperature drops are not as steep as one would expect along "wet" adiabats (15).

If such polybaric models prove physically plausible, then remelting a differentiated source at great depths would imply an early "magma ocean" that involved nearly all of the Moon.

REFERENCES

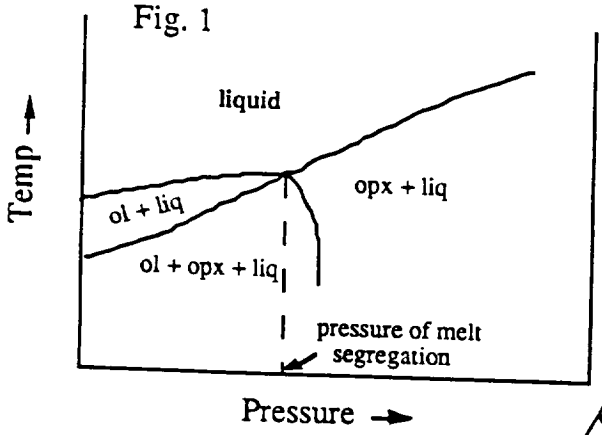
(1) Delano J.W. (1986) *J. Geophys. Res. 91*, D201-D213. (2) Green D.H., Ware N.G., Hibberson W.O., and Major A. (1971) *Earth Planet. Sci. Lett. 13*, 85-96. (3) Walker D., Longhi J., Stolper E.M., Grove T.L., and Hays J.F. (1976) *Geochim. Cosmochim. Acta 39*, 1219-1235. (4) Delano J.W. (1980) *Proc. Lunar Planet. Sci. Conf. 11th*, p. 251-288. (5) McKenzie D. (1984) J. Petrol. 25, 713-765. (6) McKenzie D. (1985) *Earth Planet. Sci. Lett. 74*, 81-91. (7) McKenzie D. and Richter F. M. (1984) *J. Geol. 92*, 729-740. (7) Klein E.M. and Langmuir C.H. (1987) *J. Geophys. Res. 92*, 8089-8115. (8) McKenzie D. and Bickle M.J. (1988) *J. Petrol. 29*, 625-679. (10) Solomon S.C. and Head J.W. (1979) *J. Geophys. Res. 84*, 1667-1682. (11) Falloon T.J. and Green D.H. (1988) *J. Petrol.* Special Lithosphere Issue, 379-414. (12) Ahearn J.L. and Turcotte D.L. (1979) *Earth Planet. Sci. Lett. 45*, 115-122. (13) Taylor S.R. Planetary Science: A Lunar Perspective, The Lunar and Planetary Institute, Houston. (14) Scott D.R. and Stevenson D.J. (1986) *J. Geophys. Res. 91*, 92383-9296. (15) Ribe N.M. (1985) *Earth Planet. Sci. Lett. 73*, 361-376. (16) Longhi J. and Pan V. (1988) *Proc. Lunar Planet. Sci. Conf. 18th*, p. 459-470.

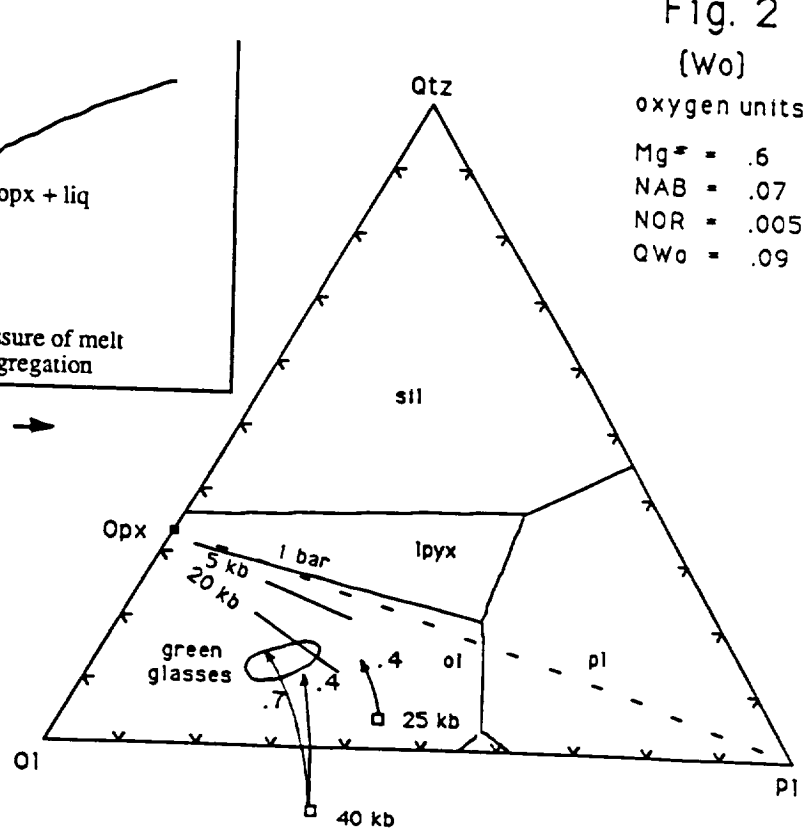
TABLE 1 (wt%)

	SiO2	TiO2	Al2O3	Cr2O3	FeO	MgO	MnO	CaO	K2O	Na2O	Mg'
model mantle	44.2	0.26	1.87	0.70	15.6	35.1	0.17	2.05	0.003	0.03	0.80
pooled melt 40 kb /0.4	43.9	1.44	7.73	0.59	20.1	15.1	0.25	10.7	0.02	0.18	0.57
Ap 14 green glass (1)	44.1	0.97	6.71	0.56	23.1	16.6	0.28	7.94	0.05	0.19	0.56

FIGURE CAPTIONS

Fig. 1. Schematic high-pressure melting relations of low-Ti mare compositions. Fig. 2. Comparison of green glass compositions (1) and calculated polybaric, pooled melt compositions. Phase boundaries are appropriate for low pressure and compositonal paprameters shown in upper right: Mg' = MgO/(MgO+FeO); NAB and NOR are albite and orthoclase fractions of the normative feldspar; QWo is the Wo coordinate of the Ol-Pl-Wo-Qtz system in mole units (15). Curved lines describe trace of cumulative pooled melt compositions. Numbers give the percentage of generated melt that is extracted in each cycle.





PRESSURE EFFECTS IN ANORTHOSITIC AND RELATED MAGMAS. J. Longhi[1], M. S. Fram[1], J. Vander Auwera[2], and J. Montieth[3]. 1: Lamont-Doherty Geological Observatory, Palisades, NY 10964 USA; 2: L. A. Gielogie-Petrologie-Geochimie, Universite de Liege, Liege B-4000, Belgium; 3: Dept. Geological Sciences, S. U. N. Y., Binghamton, NY13902 USA

Here we discuss some new and existing (1) experimental data that show the effects of pressure on crystal/liquid partition of major elements for orthopyroxene and plagioclase. Although the data are directly relevant to terrestrial massif anorthosites, they also have a bearing on lunar ferroan anorthosites.

We have run melting experiments on three starting materials in pre-saturated Pt-wire loops at QFM at 1 bar and in graphite capsules at 5 to 27 kbar. One material (500B) is rock powder from an anorthositic dike from the Nain Complex proposed to be a suspension of ~30% plagioclase in parental magma (2); the second material (HLCA) is partially crystallized synthetic glass with the composition of the average ferroan high-Al gabbro proposed to be the parental magma of the Harp Lake Complex of Labrador (3); the third (TJ) is a rock powder from the fine-grained monzonoritic border facies of the Hidra Complex of Norway (4, #80123).

Some of the results on 500B and HLCA have been reported by (1) and show: a) that even a suspension of 30% plagioclase with the 500B composition would have an improbably high liquidus temperature; b) that experimental crystallization orders for 500B and related anorthositic dikes do not agree with those observed in the dikes themselves, indicating that the dikes are heterogeneous and crystallized as open systems; c) that liquidus plagioclase (plag) in both compositions becomes progressively more sodic with increasing pressure; d) that sub-liquidus orthopyroxene (opx) in HLCA becomes progressively more aluminous with increasing pressure; and e) at 10-12 kbar, where plag, opx, and augite are near liquidus phases in HLCA, opx has the composition of the most aluminous megacrysts from the complex (Fig. 1a) and plag has the same composition as the plag in the anorthositic rocks. These features are consistent with a polybaric origin for massif anorthosites with most of the plagioclase and the few relic pyroxene megacrysts first crystallizing in the lower crust from a magma similar in composition to HLCA and then intruding upwards in plagioclase-rich mushes or suspensions (5).

Results of new experiments show that the pressure-dependence of Al_2O_3 concentration in orthopyroxene in HLCA is reversible (Fig. 1b and c). Solute rejection calculations based on the model of (6) show that, if high Al_2O_3 concentrations in opx are achieved by rapid crystal growth (e.g., 7), the expected correlation of compatible elements in opx, such as Cr, with Al is negative -- contrary to what is observed in nature (Fig. 2). Delayed nucleation of plagioclase (7) is another way to achieve unusally high concentrations of Al_2O_3 in opx that would produce a negative correlation of Cr and Al. Thus high-pressure crystallization and not kinetics is the likely explanation for the high Al_2O_3 concentrations in the opx megacrysts.

Previously, (8) demonstrated a temperature-corrected pressure effect on plagioclase composition for HLCA and 500B to 20kb. Fig 3a illustrates this effect for an additional composition (TJ) and to higher pressure (27 kbar). The vertical axis is the difference between the albite (Ab) content of plagioclase calculated at 1 bar according to the model of (9) and the albite content observed in the experiments. To a first approximation there is a composition-independent increase of ~ 1 to 2 % Ab per kbar. However, examination of the individual partition coefficients for albite and anorthite (An) versus P/T (Fig. 3b and c) shows not only clear compositional dependences, but opposite sense for pressure-dependence of K_{Ab} and K_{An}. The variation of K_{Or} versus P/T is similar to that of K_{Ab}. Despite the low pressure/depth gradient on the Moon, these data may have a bearing on the petrogenesis of lunar ferroan anorthosites (LFA), which show at least three slightly different trends on the familiar Mg' vs An diagram (10). Although differing parental magma composition is the most probable explanation, crystallization at different depths at the base of a growing crust would produce offsets in fractionation paths, as would polybaric processes such as have been proposed for massif anorthosites (5).

Fig. 4 illustrates a final aspect of the data, viz. that the simple molar partition coefficient for MgO between plagioclase and liquid has a negligible pressure dependence. Data for FeO are not shown because the analyses of mantles and rims of plagioclase crystals are prone to secondary fluorescence errors (11). These data have the most direct bearing on lunar ferroan anorthosites (LFA) where low concentrations of Fe and Mg have prompted speculations about pressure effects (11). Even though there are small compositionally dependent differences between lunar and terrestrial partition coefficients, the sense and magnitude of the pressure effect on the partition coefficient is very likely to be similar in both case. Given that most LFA appear to have crystallized from cotectic magmas (12), the low concentrations of Fe and Mg imply subsolidus exsolution of the Fe-Mg component (e.g., 13).

REFERENCES

(1) Fram M.S. and Longhi J. (1992) *Am. Mineral.*, in press. (2) Wiebe, R.A. (1990) *Am. Mineral.*, *75*. 1-12. (3) Emslie, R.F. (1980) *Geol. Surv. Can. Bull. 293*, 1-136. (4) Duschesne J.C., Wilmart E., Demaiffe D. and Hertogen J. (1989) *Precambrian Res.*, *45*, 111-128. (5) Longhi, J. and Ashwal L.D.(1984) *Proc. Lunar Planet.*

VOLATILES IN SNC PETROGENESIS: A SR SIGNAL?

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A recent experimental study (1) has confirmed a significant concentration of H_2O (~1 wt%) in the parent magma of Chassigny. This work was based upon the presence of a hydrous amphibole in a silicate melt inclusion and suggests that similar inclusions in Shergotty (2) record significant magmatic H2O concentrations as well. Speculations on the unusually high normative wollastonite component (high Ca/Al) calculated for the Nakhla parent magma (3,4) point to a role for CO_2 as well in the genesis of the nakhlite magmas (4,5). It should be instructive, therefore, to compare the compositions of the SNC parent magmas with basaltic rocks from other planets to see if there are any similarities between the SNC magmas and those for whom fluids play an important petrogenetic role. A previous survey of the non-volatile incompatible element patterns of eucritic, lunar, and terrestrial basalts showed that the SNC had patterns of rare earth (REE) and high-field strength (HFSE) element abundances distinct from any other common solar system basalts, but sharing some interesting parallels with volcanic rocks from terrestrial convergent plate margins (5), for which a petrogenetic role for fluids is generally accepted. One element omitted from that study that does have a significant susceptibility to fluid transport is Sr. So the present work attempts to fill in that gap.

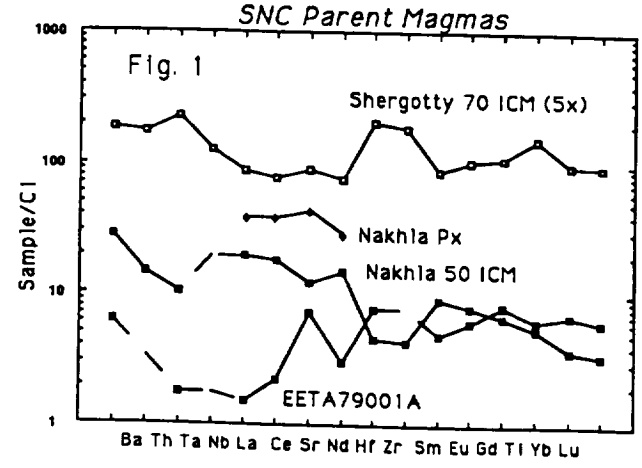
Fig. 1 illustrates the measured (6) and calculated (5,7) concentrations of the non-volatile incompatible elements in the SNC parent magmas plotted in order of decreasing incompatibility (left to right). The order is taken from (8). Previous work pointed out the generally reciprocal patterns of the Nakhla and 79001A patterns both in terms of REE and HFSE. Unlike the Antarctic shergottite, 79001A, the Shergotty parent is only weakly depleted in the light REE, but nevertheless has positive anomalies of the HFSE. The similarity of the major element compositions of Shergotty and 79001A (5), coupled with differences in the light REE and Nd-isotope ratios (9), suggest a crustal(?) assimilation event for Shergotty (5,9). There is a prominent positive Sr anomaly for 79001A relative to the adjacent REE; this pattern is based on a direct chemical analysis (6) and is, therefore, presumably real. The calculated Shergotty pattern shows a negligible Sr anomaly. The Nakhla pattern has either a positive or negative Sr anomaly depending on the type of calculation: the 50 ICM pattern is calculated from the bulk rock composition and an inferred amount of trapped intercumulus liquid (5); the Px pattern is calculated directly from the composition of a separate of augite cores (10). The ICM calculation presumes that the rock solidified as a closed system; the Px calculation presumes that the separate is pure. *In situ* measurements of Sr in the augite cores need to be made, but given that plagioclase was a late crystallizing phase in Nakhla (3), the Px calculation seems less likely to be in error. Therefore, a small positive Sr anomaly is inferred for the Nakhla parent magma.

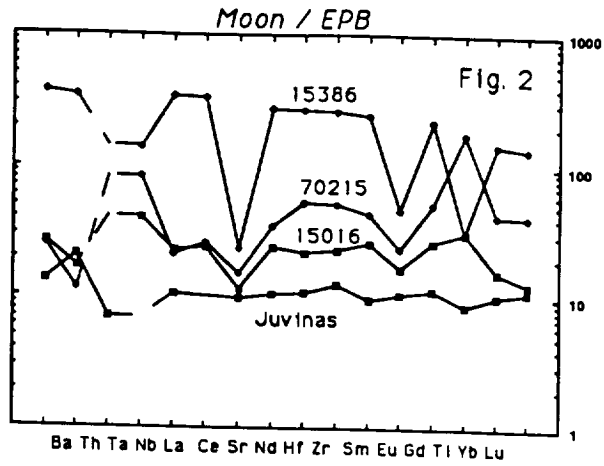
Examination of the incompatible element patterns (including Sr) of solar system basalts in whose petrogenesis volatiles play a minor or negligible role (lunar, eucritic, MORB, basaltic komatiites, continental flood basalts) shows that there are prominent Sr anomalies only for lunar basalts (Fig. 2) and these anomalies are paralleled by Eu anomalies indicating remelting of a source region depleted by the formation of an early plagioclase-rich crust (e.g, 11). Juvinas lacks a Sr anomaly (Fig. 2), as do common depleted (MORB I) and enriched (MORB II) mid-ocean ridge basalts (Fig. 3a) and typical basaltic komatiites (BK II, also Fig. 3a). The most typical flood basalts (FB D[eccan], Fig . 3b) show weak Sr anomalies, but lack the prominent HFSE/REE fractionations of the SNC. Prominent Ta-Nb anomalies are present in the less common picrites associated with flood basalts, but not the strong Hf-Zr anomalies characteristic of the SNC. Strong positive Sr-anomalies are common in ocean island basalts (OIB, Fig. 3a), in island arc basalts (IAB, Fig. 3c), boninites (Bn, Fig. 3c), and high-Al basalts (HAB, Fig. 3c). These anomalies are present even in aphyric lavas (e.g., 12). Fluids may play a role in the low degrees of melting typically inferred for OIB petrogenesis (13), however, these lavas do not have the strong Hf-Zr anomalies of the SNC. On the other hand not only do the convergent margin lavas have strong Hf-Zr anomalies, but these anomalies are both positive and negative with the positive anomalies being found in rocks with lower REE concentrations and negative Hf-Zr anomalies in rocks with higher REE. The analogy between the SNC and convergent plate margin lavas breaks down when the Ta-Nb anomalies are considered (compare Figs. 1 and 3c), however, we should expect an exact analogy because there is no evidence of plate tectonics on Mars. Nonetheless the variations in the concentrations of REE, HFSE, and Sr in convergent margin lavas are frequently attributed to fluid transport, with Sr being more mobile than the REE, which in turn are more mobile than the HFSE. Although the planet-wide differentiation on Mars is certain to be different than that observed in terrestrial convergent plate margins, it seems worthwhile to anticipate an important role for fluids in martian petrogenesis.

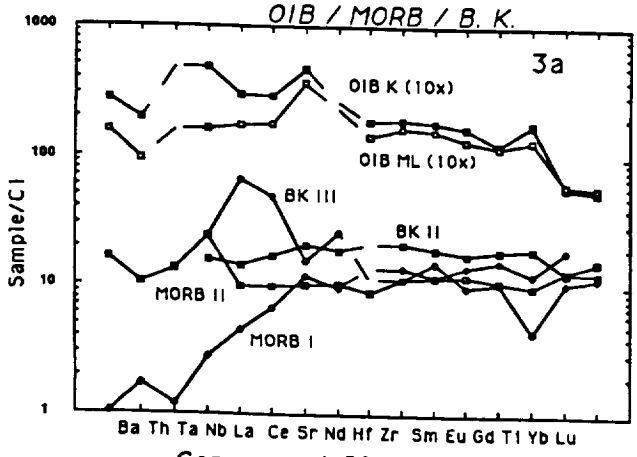
REFERENCES

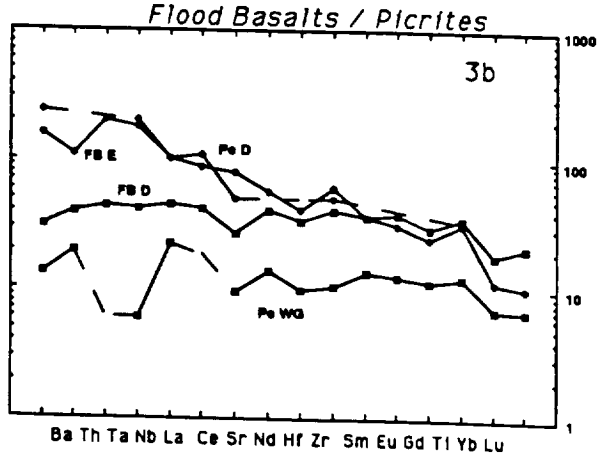
(1) Johnson M.C., Rutherford M.C., and Hess P.C. (1991) *Geochim. Cosmochim. Acta 55*, 349-366. (2) Treiman A.H. (1985) *Meteoritics 20*, 229-243. (3) Treiman A.H. (1986) *Geochim. Cosmochim. Acta 50*, 1061-

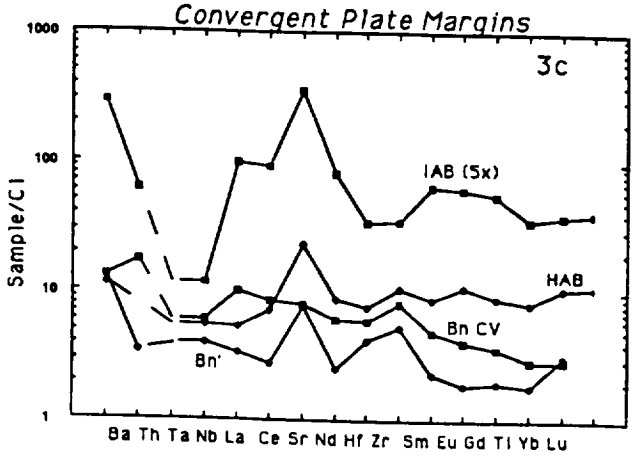
1070. 4) Longhi J. and Pan V. (1989) The parent magmas of the SNC meteorites. *Proc. Lunar Planet. Sci. Conf. 19*, 451—464.(5) Longhi (1991) *Proc. Lunar Planet. Sci. Conf. 21*, 695-709. (6) Burghele A., Dreibus G., Palme H., Rammensee W., Spettel B., Weckwerth G., and Wänke H. (1983) In *Lunar and Planetary Science XIV*, The Lunar and Planetary Institute, Houston, pp. 80-81. (7) Lundberg L. L., Crozaz G., McKay G. A. and Zinner E. (1988) *Geochim. Cosmochim. Acta, 52*, 2147—2164. (8) Wood D.A. (1979) *Geology 7*, 499-503. (9) Jones J. H. (1989)*Proc. Lunar Planet. Sci. Conf. 19*, 465—474. (10) Nakamura N., Unruh D. M., Tatsumoto M., and Hutchinson R. (1982) *Geochim. Cosmochim. Acta, 46*, 1555—1573. (11) Taylor S. R. (1982) *Planetary Science: A Lunar Perspective*, The Lunar and Planetary Institute, Houston, 481 pp. (12) Bartels K.S., Kinzler R.J. and Grove T.L. (1991) *Contrib. Mineral. Petrol. 108*, 253-270. (13) Chen C. -Y. and Frey F. A. (1985) *J. Geophys. Res., 90*, 8743—8768.











Figures: Chondrite-normalized patterns of incompatible elements arranged in order of decreasing incompatibility from left to right (8). Dashed lines indicate estimated concentrations. 1. Nakhla patterns explained in text; data from sources in (5). 2. Data from sources in (5). 3. data from sources in (5) except HAB from (12).

Conjuring Up the Red Planet

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Conjuring Up the Red Planet

by John Longhi

Orbiting spacecraft and the Viking landers have provided us with a vast amount of new data about Mars, but our understanding of Mars remains tantalizngly incomplete. We know the composition of Mars' atmosphere and soils; we know that Mars has a high-density core; and we have a huge photographic catalog of martian landforms, showing vast cratered terrains, extensive lava plains, and gigantic shield volcanoes; but we know very little about Mars' interior: the size and state of the core, the structure of the mantle, or the composition of its magmas. Future lander and sample return missions will eventually fill in many of the gaps in our knowledge. In the meantime, however, many planetary scientists believe that a small group of meteorites, called the SNC meteorites, are actually samples of Mars. If so, they permit us to make some interesting speculations.

The SNC (an acronym for three classes of meteorites: shergottites-nakhlites-chassignites) are distinct among the general class of meteorites known as basaltic achondrites. Several have traces of water-bearing minerals and the iron-oxide mineral magnetite is present indicating that their oxidation states are similar to terrestrial basalts, yet their oxygen isotope compositions show that they are definitely extra-terrestrial. Widespread speculation that the SNC meteorites came from Mars began in the late '70s after radiometric age dating showed that their minerals had crystallized no more than 1.3 billion years ago—such "recent" igneous activity being likely only on a planet large enough to retain its heat. Our much smaller Moon, for example, stopped erupting magmas about 2 billion years ago. More specific evidence came in the early '80s when analyses of neon and argon gases trapped in an impact-melted glass in one of the shergottites revealed a component of trapped martian atmosphere.[1] Since then researchers have put considerable effort into milking as many inferences about Mars as possible from the SNC meteorites.

Composition of the Surface and Interior. Measurements by the Viking Landers showed the surface of martian volcanic plains to be covered in a soil that is low in alumina (Al_2O_3) and high in iron oxide (FeO) relative to terrestrial basalts (**Fig. 1**). A direct comparison with the SNC compositions is not advisable for many elements because most of the SNCs are coarse-grained rocks that may have gained or lost crystals during slow solidification; so, unlike fine-grained basalts, bulk chemical analyses do not recover their parent magma compositions. Fortunately, a large data base derived from melting experiments allows us to estimate the composition of the parent magmas from the mineral compositions measured in the meteorites. Estimates of the parent magma compositions of the SNCs show that martian lavas

LIQUIDUS EQUILIBRIA OF LUNAR ANALOGS AT HIGH PRESSURE.

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Melting experiments have been performed in the range of 20 to 40 kbar on partially crystallized synthetic glasses in order to test the accuracy of the polybaric fractional fusion model for picritic lunar green glasses (1). Results show that the model predicts the position of the olivine(ol) + orthopyroxene(opx) liquidus boundary within the uncertainty of the measurements, but that details of the calculations are subject to change because of new crystal/liquid partitioning data for olivine and pyroxene.

Volcanic glasses with picritic compositions (2) have posed a thorny problem for petrologists: their compositions are multiply-saturated with olivine and orthopyroxene at pressures of 17.5 to 25 kbar (cf. 3) and traditionally the pressure of multiple saturation is interpreted as the minimum pressure of melt segregation, yet on the Moon these pressures translate to depths of 350 to more than 500 km.. The physics of transporting melt these distances through cracks without altering the melt composition have been modeled (4) and the possibility of such transport seems remote at a time in lunar history when the thickness of the elastic lithosphere was on the order of 50 to 100 km (5). Thus polybaric melting models analogous to those proposed for terrestrial midocean ridge basalts (6) provide attractive alternatives. Because of the paucity of data, algorithms in the polybaric melting model proposed by the author (1) depended upon several extraoplations of phase equilibria and crystal/liquid partitioning from experiments at lower pressures on terrestrial compositions, so a series of melting experiments was initiated to provide data at high pressure with which to test and refine the model.

Partially crystallized glasses were synthesized with compositions designed to produce multiply-saturated liquids at high pressure and run in high purity iron capsules for 1 to 2 days in a piston-cylinder apparatus. The resulting charges contained remarkably homogeneous olivine + pyroxene ± garnet set in fine intergowths of quench crystals. In one charge a comparison of the average of 36 spot analyses (electron microprobe) collected on a grid was compared with the average of 4 rastered-beam analyses (100 μm on a side): all elements agreed within the standard deviations indicating that raster analyses of highly mafic quench intergrowths (no plagioclase) are accurate. Fig. 1 shows the compositions of melts in equilibrium with ol and opx (open squares) together with segments of the ol + opx liquidus boundary calculated for each melt according to the algorithms in (1). In each case except at 25 kbar the melt composition has a slightly higher Qtz component than the predicted liquidus boundary, however, the differences are not significant with respect to the overall uncertainty of the model.

The new experiments also provide relevant data on partitioning and phase equilibria. There is a dramatic increase in the molar partition coefficient (D*) for Al_2O_3 between opx and liquid(liq) from an average value of ~0.1 at low pressure to a value of 0.5 at 30 to 40 kbar. Values of D* for TiO_2 and Cr_2O_3, however, remain constant (~0.15 and ~ 2, respectively) over this pressure range. The different behaviors can be explained in terms of liquid structure: Al is tetrahedrally coordinated in silicate liquids at low pressure, but these tetrahedral sites become increasingly unfavorable with respect to the octahedral sites available in pyroxene as pressure increases; whereas Ti and Cr find no greater advantage in pyroxene with increasing pressure because they are already in octahedral sites in the liquid at low pressure. The previous model (1) employed a single value of D* (0.35) for opx and liq at high pressure that is now obviously too low for pressures in the range of 30 to 40 kbar. A consequence of using a pressure-dependent D* for Al is that now it is more difficult to deplete the source in Al_2O_3 at high pressure and extraction of melt at each pressure must, therefore, be more efficient to produce a polybaric pooled melt with green glass composition.

Fig. 2 shows the approximate location of liquidus boundaries at 25 kbar (dashed curves) in two projections based on multiply-saturated experimental liquid compositions (squares). Crystal compositions are shown as circles. Phase equilibria at 30 kbar are similar albeit displaced slightly farther away from Qtz. It is not clear yet whether in these intermediate Mg' liquids there is a liquidus boundary in the {Ol} projection separating fields of ol + pigeonite(pig) and ol + aluminous augite(aug) or whether there is a single continuous field of ol + clinopyroxene (cpx). In any case because most estimates of lunar mantle compositions have low Wo contents, melting of the interior at P > 25 kbar probably begins in the presence of ol, opx, and pig. Similar behavior has been observed in experiments on martian analogs (7). If there is sufficient Al in the source to stabilize garnet(gar), as there is in most estimates of whole Moon composition (8), then the equilibrium at the solidus will be opx + liq = ol + gar + pig. As in the case of the martian analogs, the initial melt will be nepheline-normative.

Finally, the Fe-Mg exchange coefficient for ol and liq (K_D) has similar values in the high pressure experiments as in previously reported low pressure experiments on lunar basalts (3) at comparable concentrations of TiO_2 (cf. Fig. 3 in ref. 3). However, even though TiO_2 levels are similar, the silica contents of the high pressure liquids are much lower than those of the basalts, so that tendency for increasing pressure to

increase K_D at constant composition (9) is offset by the shift in the ol + opx liquidus boundary to lower silica with increasing pressure.

REFERENCES

(1) Longhi J. (1992) *Proc. Lunar Planet. Sci. 22, 343-353.* (2) Delano J.W. (1986) *J. Geophys. Res. 91,* D201-D213. (3) Longhi J. (1992) *Geochim. Cosmochim. Acta, 56,* 2235-2251. (4) Spera F. J. (1992) *Geochim. Cosmochim. Acta, 56,* 2253-2265. (5) Solomon S.C. and Head J.W. (1979) *J. Geophys. Res. 84,* 1667-1682. (6) Klein E.M. and Langmuir C.H. (1987) *J. Geophys. Res. 92,* 8089-8115. (7) Bertke C. M. and Holloway J. R. (1990) *Lunar Planet. Sci. XXI,* 73-74. (8) BVSP (1981) *Basaltic Volcanism on the Terrestrial Planets,* Pergamon. (9) Ulmer P. (1989) *Contrib. Mineral. Petrol., 101,* 261-273.

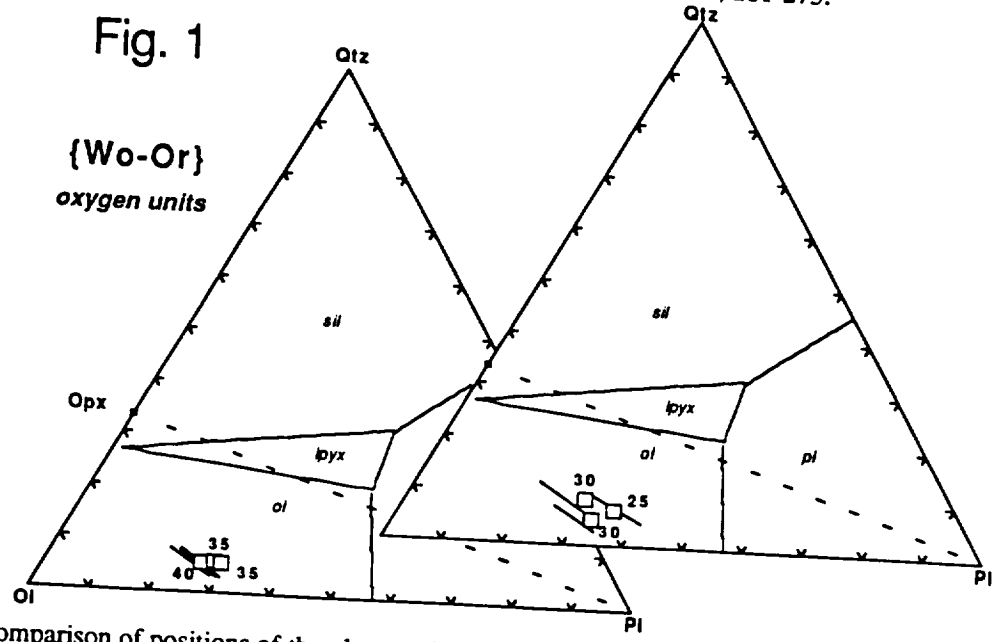

Fig.1. Comparison of positions of the ol + opx liquidus boundary in wollastonite (Wo) projection: squares are experimental determinations (this study); line segments are calculations of the liquidus boundary based on the algorithms of (1) and using the observed liquid compositions as inputs. Low pressure liquidus boundaries are shown for reference.

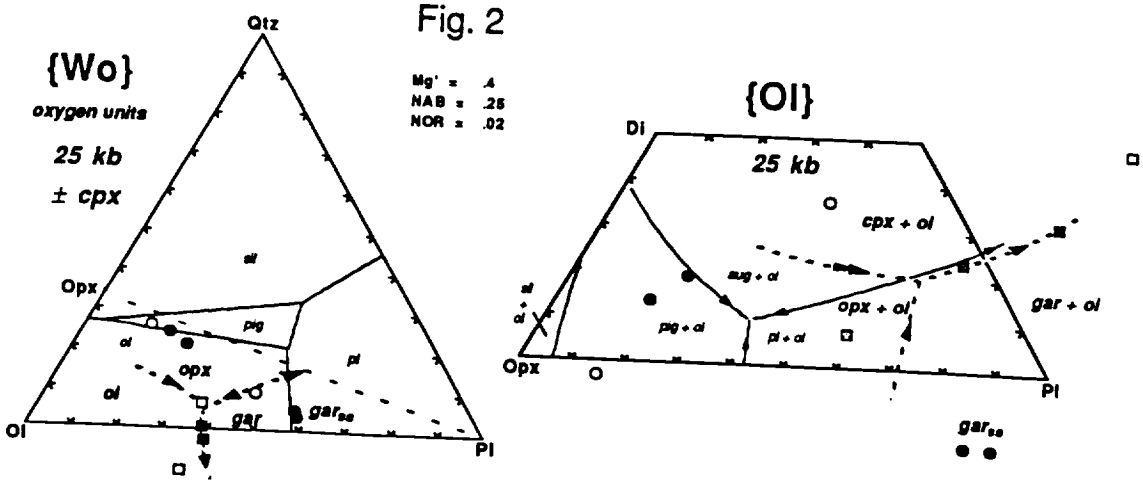

Fig. 2. Schematic liquidus equilbria at 25 kbar in Wo and Ol projections. ol-pig-gar-saturated liquids shown as solid squares; ol-opx- and ol-cpx-saturated liquids shown as open squares. Crystal compositions shown as solid or open circles as appropriate. Dashed curves are inferred 25 kbar liquidus boundaries; solid curves are 1 bar liquidus boundaries.

THE MONZONORITE-ANORTHOSITE CONNECTION: THE PETROGENESIS OF TERRESTRIAL KREEP.

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There is a suite of rocks typically associated with Proterozoic massif anorthosites that bear some interesting similarities to lunar KREEP. In many cases (e.g., Laramie [1]) these rocks are plutonic and have traditionally been referred to as the jotunite-mangerite- ±charnockite- ±syenite suite. However, in the Rogaland district of southwestern Norway, where they are referred to as "monzonorites", these rocks are also present as fine-grained dikes and as the chill margin of a layered intrusion [2,3], and thus approximate magmatic liquid compositions are readily obtained by chemical analysis. Monzonorites are typically enriched in incompatible lithophile elements such as K (alkali feldspar is present), the rare earths (REE), and P. They have intermediate to low Mg', low-Ca pyroxene, and more evolved types have low Ti/Sm ratios. Much debate has developed over attempts to explain the link between monzonorites and massif anorthosites. One feature seems clear: monzonorites and associated anorthosites have different initial isotopic ratios [3,4], so a simple relation is not possible. However, there is apparently a continuum in major elements between the monzonorites and gabbros believed to represent magmas parental to the anorthosites. This continuum suggests a link via high pressure fractionation coupled with assimilation. Although more complicated, this scenario is similar that evoked for the early Moon: following the formation of ferroan anorthosites, continued fractional crystallization of the residual liquids at the base of the crust led to the formation of KREEP [5]. An attempt is made here to establish a link between monzonorites and high-Al gabbros which are nearly always found as ancillary intrusions associated with anorthosites, and which may record processes in lower crustal magma chambers.

Fig. 1 illustrates some of the evidence for a link by fractionation between monzonorites and gabbros . The upper panels show a continuous variation in TiO_2, P_2O_5, and MgO between gabbros and ferrodiorites of the Harp Lake Complex of Labrador [6] and the monzonoritic dikes of the Rogaland district of Norway [3]. Patterns for other elements (SiO_2, Al_2O_3, FeO, etc.) versus MgO are also continuous. Interpretation of the elemental variations are not straightforward, however. The peaked TiO_2-MgO pattern is as expected for fractional crystallization of a magma that eventually becomes saturated in ilmenite. Except for the subset of analyses with P_2O_5 above 2 wt %, the P_2O_5-MgO pattern is generally similar and indicates crystallization of a phosphate (apatite) between 2 and 3 wt % MgO. The samples with the highest P_2O_5 also have lower SiO_2 (43-45 wt%) and higher FeO (15-20 wt%) than those one the main trend. These samples are from dikes that show large regional variation in composition, and it seems possible that these dikes underwent an internal differentiation caused by differential flow of a crystalline matrix and interstitial liquid. The high-P-Fe, low-Si samples thus may be partial cumulates. If so, then the highest TiO_2 concentrations may be enhanced by accumulation as well. Experimental liquids obtained from gabbroic (HLCA) and monzonoritic (TJ) starting materials tend to mimic the high-MgO portion of the natural trend although it is clear that the specific compositions employed cannot be directly related. The steep trend of increasing TiO_2 with decreasing MgO for the natural samples in Fig. 1 indicates that fractionation must have involved extensive Fe-Mg equilibration.

Trace elements provide evidence of further complications. Several of the monzonoritic dikes lack negative Eu anomalies in their REE patterns and have only weak depletions of chondrite-normalized Sr relative to the light REE [3] — hardly what is expected from liquids residual to the formation of anorthosite. Phase equilibria complicates matters still more. Experiments on one of the most primitive monzonorites (sample TJ, the chill margin of the Bjerkeim-Sokndal intrusion [7]) show that plagioclase(pl) and orthopyroxene(opx) are together on the liquidus from 10 to 13 kbar. This pressure is consistent with experiments that showed aluminous opx megacrysts found in anorthosite plutons to have formed in this pressure range [8]. However, as Fig. 2 illustrates, the TJ composition is close to a thermal divide on the pl-opx(±pigeonite)-augite(aug) liquidus boundary in this pressure range, yet it would be extremely fortuitous for a residual liquid to remain on a thermal divide. One way to generate a liquid on a thermal divide would be in a second stage fusion of the lower crust: the first stage removes a granitic component leaving a pyroxene-plagioclase residuum, which would necessarily produce melts on the thermal divide at these pressures. This residuum would likely be depleted in incompatible elements and have positive Eu and Sr anomalies with respect to the other REE, so it would be necessary to invoke small degrees of melting to produce incompatible element enrichments and flatten out the Eu and Sr anomalies. Such a scenario would not, however, readily explain the continuous variation in major elements evident in Fig. 1 nor the continuous variation in mineral compositions between anorthositic and monzonoritic rocks [1].

A final option is assimilation-fractionation. In this scenario the compositions of the monzonoritic parent magmas (derived by high pressure crystallization of gabbroic liquids) are altered by assimilation of

granitic crust at lower pressures (i.e. ≤ 7 kbar) where the pyroxene-plagioclase thermal divide is no longer present. Thus the present location of the TJ composition near the thermal divide is accidental and the incompatible element patterns are hybrids. Further work is needed to verify this hypothesis.

REFERENCES

[1] Kolker A. and Lindsley D. H. (1989) *Amer. Mineral., 74*, 307-324. [2] Duschene, J.C., Roelandts I., Demaiffe D., and Weiss D. (1985) *Contrib. Mineral. Petrol. 90*, 214-225. [3] Duchesne J.C., Wilmart E., Demaiffe D. and Hertogen J. (1989) *Precambrian Res., 45*, 111-128. [4] Geist D. J., Frost C. D., and Kolker A. (1990) *Amer. Mineral., 75*, 13-20. [5] Warren P. H. and Wasson J. T. (1979) *Rev. Geophys. Space Phys.* 17, 73-88. [6] Emslie, R.F. (1980) *Geol. Surv. Can. Bull. 293*, 1-136. [7] Duchesne J.C. and Hertogen J. (1988) *C. R. Acad. Sci. Ser. II, 306*, 45-48. [8] Fram M.S. and Longhi J. (1992) *Am. Mineral., 77*, 605-616.

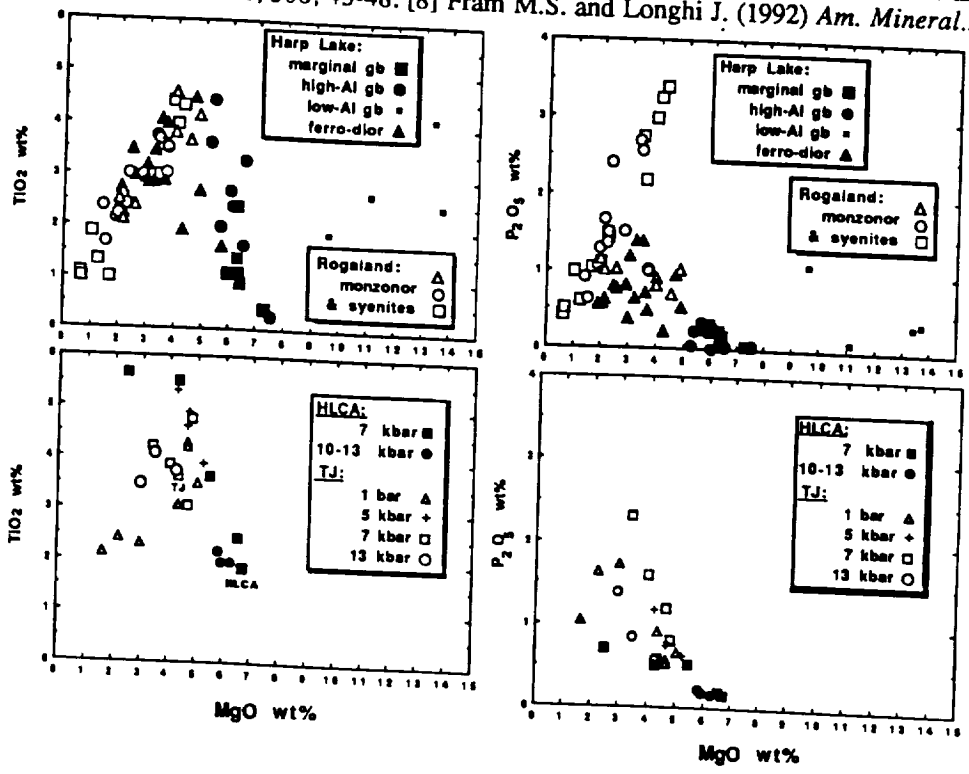

Fig. 1

Fig. 1 Major element variation in gabbros and monzonoritic rocks associated with the Harp Lake (Labrador) [6] and Rogaland (Norway) [3] anorthosites compared with liquid compositions ([8] and this study) obtained from experiments in graphite capsules on natural compositions. Fig. 2 Liquidus equilibria on the plagioclase + ilmenite saturation surface at lower crustal pressures. Liquids shown as filled squares; dashed tie lines connect coexisting pyroxene compositions (small open circles).

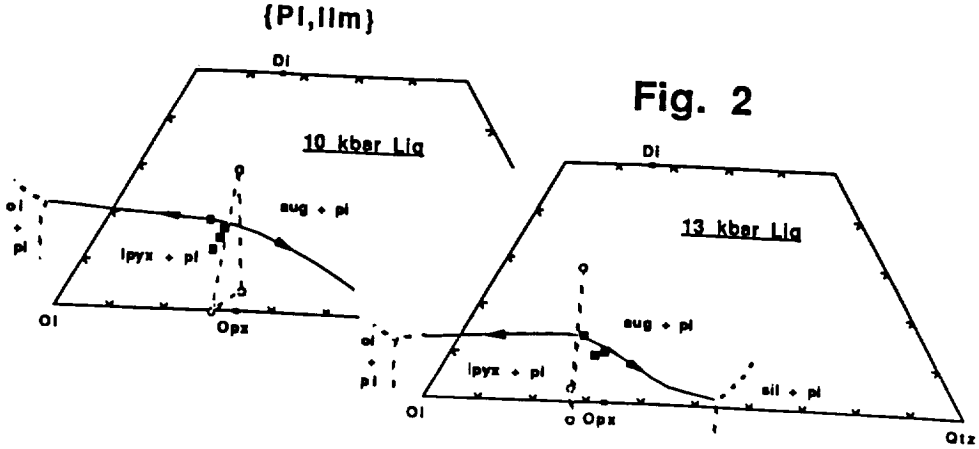

Fig. 2

LIQUIDUS EQUILIBRIA OF LUNAR ANALOGS IN THE GARNET STABILITY FIELD.

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New melting experiments have been performed in the range of 28 to 30 kbar on partially crystallized synthetic glasses in order to refine parameterizations of liquidus boundaries and crystal/liquid partition coefficients employed in polybaric fractional fusion (PFF) calculations (1,2). Results confirm that PFF of a low-Al_2O_3 (< 1.5 wt%), low-Mg' (~0.80) source may yield composite melts similar in composition to the picritic green volcanic glasses (3): a) if pressure-release melting (~1%/kbar) begins very deep within the Moon (1200 km, ~40 kbar), b) if ~60% of the available melt effectively separates after each 1-2 % melting, and c) if there is some means of aggregating the melt produced over a range of depths. New results from high Mg' compositions also agree with predictions (4) that the clinopyroxene(cpx) coexisting with olivine(ol) + orthopyroxene(opx) ± garnet(gar) on the solidus of anhydrous planetary mantles in the garnet stability field is not only low in CaO (5-8 wt%), but is also supercritical.

Volcanic glasses with picritic compositions (3) have posed a thorny problem for petrologists: their compositions are multiply-saturated with olivine and orthopyroxene at pressures of 17.5 to 25 kbar (cf. 5) and traditionally the pressure of multiple saturation is interpreted as the minimum pressure of melt segregation, yet on the Moon these pressures translate to depths of 350 to more than 500 km.. The physics of transporting melt these distances through cracks without altering the melt composition have been modeled (6) and the possibility of such transport seems remote at a time in lunar history when the thickness of the elastic lithosphere was on the order of 50 to 100 km (7). Thus polybaric melting models analogous to those proposed for terrestrial midocean ridge basalts (8) provide attractive alternatives. Algorithms in the polybaric melting model proposed by the author (1) depended upon several extrapolations of phase equilibria and crystal/liquid partitioning from experiments at lower pressures on terrestrial compositions, so a series of melting experiments was initiated to provide data at high pressure with which to test and refine the model. The first round of experiments [(2) - partial results in Fig. 1a] produced melts with Mg' (0.40-0.45) lower than that appropriate for the green glasses. The new experiments (Fig. 1b and c) have multi-saturated melts with Mg' ranging from 0.50 to 0.78 — the lower end of the range are more appropriate to green glass source regions, whereas the higher values are appropriate to proposed bulk Moon compositions that are similar to the Earth's upper mantle (e.g., 9)

The new experiments also provide relevant data on partitioning and phase equilibria. There are dramatic increases in the molar partition coefficients (D*) for Al_2O_3 for opx/liq and aug/liq from average values of ~0.1 and ~0.2, respectively, at low pressure to values of ~0.5 and ~0.7 at 30 to 40 kbar (Fig. 2). Although in both cases the strongest correlating factor is P/T, the aug/liq coefficient has several statistically significant compositional dependences including the wollastonite fraction in pyroxene (WoPx). So an independent expression for WoPx, appropriate for augite at low pressures and supercritical cpx at high pressures, was obtained by linear regression (Fig. 2). The original model (1) employed constant values of D* for opx/liq (0.35) and cpx/liq (0.44) at high pressure that are now obviously too low for pressures in the range of 30 to 40 kbar. A consequence of using a pressure-dependent D* for Al is that now it is more difficult to deplete the source in Al_2O_3 at high pressure and extraction of melt at each pressure must, therefore, be more efficient to produce a polybaric pooled melt with green glass composition. Consequently, permissible melt extraction efficiency is now more than 60% as compared to 40% originally.

Fig. 1 shows the approximate location of liquidus boundaries (dashed curves) projected from the olivine component at 30 kbar for low (a) and intermediate Mg' liquids (b) and at 28 kbar for high Mg' liquids (c). Experimental liquid compositions are shown as squares, crystal compositions as circles. Filled symbols indicate olivine and liquid coexisting with two other solid phases, open symbols indicate one other solid phase. Fig. 3a contrasts liquidus boundaries at 1 bar (light solid lines) and 30 kbar (dashed lines) for Mg' appropriate for mare basalts. Opx is not stable at 1 bar, but has a relatively large stability field at 30 kbar; the cpx that coexists with opx, gar, and ol is constrained to be low in CaO; minimum melts of garnet peridotite assemblages are tholeiitic; and the psuedoinvariant point is a peritectic with opx in reaction with liquid. The 30 kbar topology in Fig. 3b, which is appropriate to picritic compositions, is similar to that in 3a: the data here fix the position of the ol + opx + cpx liquidus boundary and definitively show opx in reaction with liquid. The bulk composition of the green glass source region used in the polybaric melting calculations (1) is shown as a cross. Because the cross is lower in Al than the pyroxenes coexisting with gar and liq, garnet is not stable on the solidus. The coverage of data in Figs. 3a and b is ambiguous as to whether both low- and high-Ca pyx are stable as separate phases. However, the data in Fig. 3c show that cpx solid solution is continuous in petrologically relevant compositions as predicted by Bertka and Holloway (4). Liquidus boundaries in Fig. 3c are shifted to higher Al than those in Figs. 3a and b. This shift is the result of higher Mg' and alkalies. The projection of primitive terrestrial upper mantle compositions (PUM - depleted and undepleted (9)) shows that

garnet is on the solidus of terrestrial-like bulk Moon compositions (8). The considerable difference in Al contents between the PUM and model green glass source regions is not likely to be the result of differing degrees of basalt extraction — pyroxene accumulation seems to be essential to produce the low Al contents of the green glass source.

REFERENCES

(1) Longhi J. (1992) *Proc. Lunar Planet. Sci. 22*, 343-353. (2) Longhi J. (1993) *Lunar Planet. Sci. XXIV*, 895-896. (3) Delano J.W. (1986) *J. Geophys. Res. 91*, D201-D213. (4) Bertke C. M. and Holloway J. R. (1993) *J. Geophys. Res. 98*, 19,755-19,767. (5) Longhi J. (1992) *Geochim. Cosmochim. Acta, 56*, 2235-2251. (6) Spera F. J. (1992) *Geochim. Cosmochim. Acta, 56*, 2253-2265. (7) Solomon S.C. and Head J.W. (1979) *J. Geophys. Res. 84*, 1667-1682. (8) Klein E.M. and Langmuir C.H. (1987) *J. Geophys. Res. 92*, 8089-8115. (9) Warren P. W. (1986) in *Origin of the Moon*, pp.279-310, The Lunar and Planetary Institute, Houston.

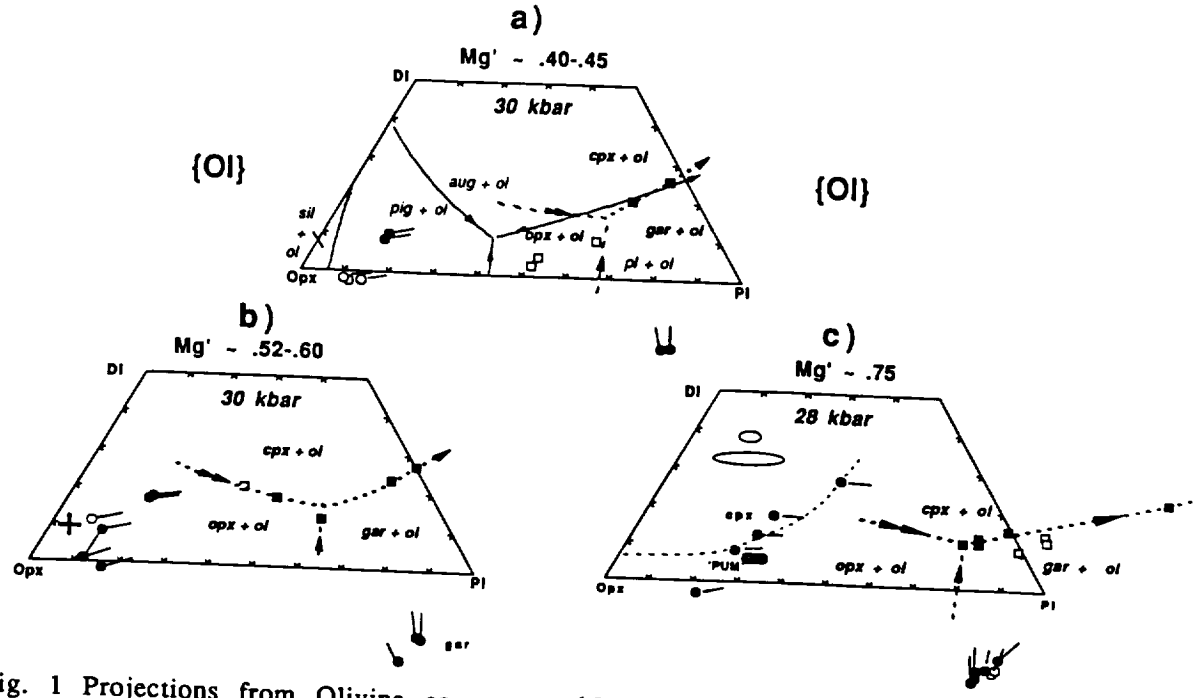

Fig. 1 Projections from Olivine component {Ol} onto a portion of the Wollastonite(Wo) - Orthopyroxene(Opx) - Plagioclase(Pl) plane. Symbols described in text. Arrows show direction of decreasing temperature; double arrows show reaction boundaries. Olivine coexists with all phases: a) fo_{67-72}, b) fo_{75-80}, c) fo_{87-91}. Large ellipses in c) show range in composition of liquid analyses (~30 micron raster of quench crystals); smaller ellipses show compositional ranges of pyroxene analyses (spot).

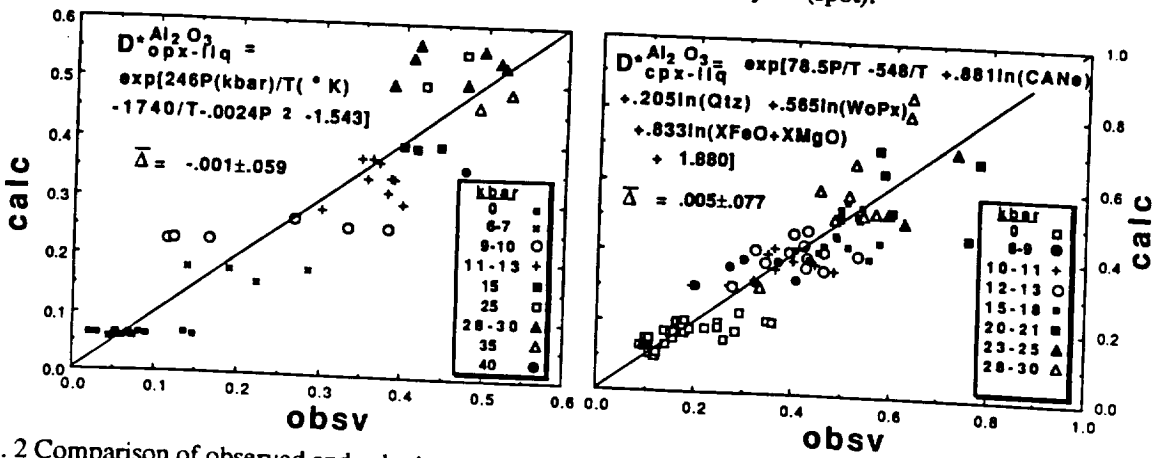

Fig. 2 Comparison of observed and calculated molar Al_2O_3 partition coefficients (D*) for aug/liq and opx/liq.

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Silicate Liquid Immiscibility in Isothermal Crystallization Experiments

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Isothermal crystallization experiments involving various mixtures and configurations of synthetic two-liquid pairs in iron capsules sealed in evacuated silica glass tubes show incompatibility under some conditions, compositional convergence under other run conditions, and further unmixing under still others. Results show that natural silicate liquid immiscibility may coexist stably with plagioclase, contrary to what is observed in simple systems. Comparison of results of isothermal and controlled-cooling-rate experiments on similar compositions suggests that kinetic barriers may have prevented the crystallization of silica minerals in the controlled-cooling-rate experiments. Multiple regression of the observed two-liquid separations against various chemical parameters indicates that P_2O_5 concentration, *Mg'*, and the plagioclase and olivine contents of the mafic liquid have significant effects on unmixing. Calculation of two-liquid separation in model melts calculated to be residual to ferroan anorthosites and presumed to be parental to *ur*KREEP indicates that silicate liquid immiscibility very probably developed during the latest stages of early lunar differentiation. However, there appears to be little evidence of two-liquid fractionation in the K_2O/P_2O_5 ratio of KREEP that lies within the range of values of low-Ti mare basalts. Immiscibility may have played a role in the formation of some lunar granites, but because these rocks have mafic minerals with intermediate *Mg'*, they could not have formed as residua of the lunar "magma ocean," which was hyperferroan. Lunar granites most probably crystallized as the plutonic residua of KREEP and high-Mg magmas.

INTRODUCTION

There are long-standing proposals that silicate liquid immiscibility (SLI) played a role in the petrogenesis of lunar granitic rocks (*Hess et al.*, 1975; *Taylor et al.*, 1979). *Neal and Taylor* (1989a) repeated this proposal recently with the additional suggestion (*Neal and Taylor*, 1989b) that the liquid precursor of these granites was *ur*KREEP (*Warren and Wasson*, 1979). Given the apparent importance of lunar granites and primitive KREEP as assimilants in mare basalts (*Shervais et al.*, 1985) and in high-Mg magmas (*Longhi*, 1981; *Warren*, 1988), it seems prudent to learn as much as possible about the development and effects of SLI in lunar magmas. There already is a considerable body of work describing trace element partitioning between immiscible liquid pairs in simple (*Watson*, 1976) and natural compositional systems (*Ryerson and Hess*, 1978), and these studies are in general accord. However, the existing experimental data on major elements suggests a possible disparity in phase equilibria between the simple laboratory systems, which petrologists consult as models of the complex natural system, and the natural system itself. Furthermore, the experimental major element data on natural lunar immiscible liquids are somewhat erratic and, consequently, the bounds of the two-liquid field are difficult to parameterize (see below). These considerations prompted this study.

Figure 1 illustrates two of the relatively simple phase diagrams used in discussions of SLI and depicts some of the two-liquid pairs developed in controlled cooling rate experiments on natural and synthetic lunar compositions (*Rutherford et al.*, 1974; *Hess et al.*, 1975, 1978). The liquidus boundaries in Fig. 1a are taken from a portion of the Fe_2SiO_4 (Fa) - $KAlSi_2O_6$ (Lc) - SiO_2 (Sil) system (*Roedder*, 1951; *Visser and Koster van Groos*, 1979a). Figure 1b shows that the effect of adding $CaAl_2Si_2O_8$ (An) to the Ol (Fa) - KS join in Fig. 1a (*Irvine*, 1976) is to suppress the two-liquid field. This is a mildly disturbing circumstance given the ubiquity of plagioclase in late stage magmas. *Visser and Koster van Groos* (1979b) have shown that adding less than 1 wt.% P_2O_5 to the Fa-Lc-Sil join expands the two-liquid field sufficiently for fayalite and K-feldspar to coexist with an immiscible liquid pair. Other studies on the Fa-Lc-Sil system show that increasing pressure (*Watson and Naslund*, 1977) and increasing oxygen fugacity (*Naslund*, 1976) also expand the two-liquid field, but not to the point of reaching the K-feldspar field. These data suggest that natural levels of P_2O_5 and other high field-strength elements will expand the two-liquid field in differentiating magmas, but leave unanswered the question of whether SLI will coexist with plagioclase. The only published experimental data in which plagioclase coexists with immiscible silicate liquids are those produced in controlled cooling rate experiments. Figure 1a depicts the two-liquid pairs coexisting with plagioclase in runs on lunar rock powders and on synthetic lunar compositions that have not been doped with trace elements.

There is considerable variation in both the length and orientation of the two-liquid tielines that does not correlate with temperature or any simple compositional parameter, such as *Mg'* [molar MgO/(FeO+MgO)] or P_2O_5 concentration. Some of the scatter is no doubt due to difficulties in analyzing the characteristically small globules that develop in these experiments. However, there is an additional concern: Despite the fact that most of these compositions project into the silica field, no silica phase was reported in the run tables of

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25.

Complex Magmatic Processes on Mars: Inferences from the SNC Meteorites

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Comparison of the major-element, incompatible-trace-element, and isotopic compositions of the parent magmas of the SNC (shergottites-nakhlites-Chassigny) meteorites with eucritic, lunar, and terrestrial basalts shows that their parent body (Mars) has a unique magmatic style. The major-element and isotopic compositions of the Nakhla and Chassigny meteorites (1.3 b.y.) suggest extensive melting of source regions with much stronger time-integrated light rare earth element (REE) depletions than those observed on Earth. The incompatible-element patterns of these magmas, however, indicate extremely strong enrichments of the light rare earth elements, suggestive of miniscule degrees of partial melting. Such compositional patterns are commonly found in basalts from terrestrial intraplate ocean islands (OIB), but relative enrichments are much greater in the martian magmas. Also, the Nakhla incompatible-element pattern (chondrite-normalized) shows prominent relative depletions of high-field-strength elements (HFSE), such as Ta, Nb, Hf, and Zr, that are not typically observed in OIB but are characteristic of island arc basalts (IAB). Closer examination of the depletions of the HFSE relative to the REE indicates that the martian pattern (Ta-Nb > Hf-Zr) is actually more similar to that of terrestrial carbonatites than to IAB. Because of their low viscosity, carbonatitic magmas at the base of the melt zone would greatly facilitate the two-phase porous flow process proposed by McKenzie (1984, 1985) at melt fractions <1% and thus might also account for the strong REE fractionation when mixed at higher levels with more extensive silicate melts. Furthermore, CO_2, which expands the orthopyroxene liquidus field at high pressure, might account for the unusually high estimates of CaO in the Nakhla parent magma. The shergottites have incompatible-element patterns—depletion of the light-REE, positive-HFSE/REE anomalies—complementary to Nakhla that suggest they were generated from a source depleted in a Nakhla-like component. This complementary relation is consistent with a younger age for the shergottites such as 180 m.y. (Jones, 1986). If this young age is correct, then the Antarctic shergottites were generated by a Sm/Nd-enrichment process (fractional fusion?) and came to the martian surface relatively uncontaminated. The classic shergottites (Shergotty and Zagami), however, also assimilated a long-term light-REE-enriched component that presumably is ancient crust. If this crust formed at ~4.4 b.y. and was complementary to the Nakhla source, then it had $\epsilon_{Nd} \sim -20$ at 180 m.y. ago. Also, this crust would have Sm/Nd ~ 0.15, which is well within the range of enriched basalts, which in turn suggests that if primordial melting were volatile free or dominated by CO_2, the ancient martian crust would be composed of transitional to alkali basalts, whereas if the ancient melts were H_2O-rich, the ancient crust would be andesitic.

INTRODUCTION

The SNC meteorites (shergottites-nakhlites-Chassigny), which have traces of hydrous minerals and terrestrial-like feldspars and oxidation state (*Stolper et al.*, 1979), form a distinct petrologic group among the basaltic achondrites. Their parent body, however, formed from an oxygen isotopic reservoir distinct from the Earth and the other meteorites (*Clayton and Mayeda*, 1983). A number of lines of evidence point to a martian origin for the SNC meteorites (*Wood and Ashwal*, 1981; *McSween*, 1985). These include relatively young crystallization ages of 1.3×10^9 or less (*Nakamura et al.*, 1982a,b; *Shih et al.*, 1982), implying derivation on a relatively large planetary object (*Walker et al.*, 1979); the discovery of weakly shocked lunar meteorites in Antarctica (*Marvin*, 1983), proving that impacts can eject rocks from planetary objects without obliterating their texture and mineralogy; the general similarity of the compositions of the martian soil and the Shergotty meteorite (*McSween and Stolper*, 1980); and the discovery of trapped martian atmosphere in one of the Antarctic shergottites (*Bogard and Johnson*, 1983). Although the SNCs are isolated samples without any stratigraphic or tectonic context, there is a wide range of petrological and geochemical data that allow inferences about the magmatic processes that produced them as well as the composition and history of their source regions. This paper draws on these data as well as parallel data from lunar, terrestrial, and other meteoritic basalts in order to reconstruct as much as possible of the magmatic processes and history of Mars.

MAJOR ELEMENTS

One of the difficulties encountered in the study of the SNC meteorites is the fact that most of them are igneous cumulates or partial cumulates (*Stolper and McSween*, 1979). Thus their bulk compositions, which can be measured directly, are not the same as the compositions of the magmatic liquids from which they crystallized. It is these parent liquid compositions upon which inferences about magmatic processes and source regions must be based. *Longhi and Pan* (1989) have calculated the parent magma compositions for several SNC meteorites. Their compositions along with other calculations taken from the literature are shown in Fig. 1 and discussed briefly below.

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Comparative liquidus equilibria of hypersthene-normative basalts at low pressure

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ABSTRACT

The low-pressure liquidus boundaries of natural basaltic liquids have been parameterized as quasi-linear empirical functions of projection coordinates in the Ol-Pl-Wo-Qtz model system. Secondary variables are simple chemical parameters such as Mg′ and the albite and orthoclase fractions of the normative feldspar. These empirical expressions lend themselves to rapid construction of liquidus diagrams as well as forming an integral part of quantitative models of fractional and equilibrium crystallization. The position and reaction relations of the liquidus boundary between olivine (ol) and low-Ca pyroxene (lpyx) are strongly dependent on Mg′ such that olivine reacts with magnesian liquids along the boundary, but precipitates from more ferroan liquids. The transition from a reaction to a crystallization relationship in liquids with intermediate Mg′ is complex, and it is possible for olivine to precipitate during equilibrium crystallization and react during fractional crystallization of the same liquid on the ol-lpyx boundary. A survey of the silicate-liquidus equilibria appropriate to mare, midocean ridge, and calc-alkaline basalts shows that the different alkali contents of these magma types are responsible for the different proportions of mafic and feldspathic components observed in cotectic liquids of each type. Different alkali contents can even produce important differences, such as silica enrichment vs. silica depletion, in the liquid lines of descent (LLD) of parent liquids that have the same initial Mg′ and crystallization sequence. Liquid lines of descent are in general different from the equilibrium crystallization paths that petrologists have employed to construct liquidus diagrams. If one were to use only the plagioclase-component projection, there is the potential for confusing a low-pressure LLD with a high-pressure trend.

INTRODUCTION

Predicting the liquidus equilibria of basaltic magmas has been a goal of petrologists ever since the first melting experiments. Among the first credible attempts to predict liquidus equilibria of natural compositions were those of O'Hara (1968) and Irvine (1970), who employed the bulk compositions and experimentally determined crystallization orders of basalts to construct a series of pseudoternary liquidus diagrams. Walker et al. (1972) carried this graphical approach a step farther by using microprobe analyses of experimentally produced multisaturated liquids to fix the positions of the liquidus boundaries (Fig. 1, top). As recognized by these authors, these diagrams were reliable only for compositions similar to those on which the diagrams were based. Roeder (1975) introduced a second approach that lent itself more readily to calculating fractional crystallization paths. He derived analytical temperature-composition expressions for the olivine and plagioclase liquidus surfaces via multiple linear regression. With these expressions it became possible to compare the calculated liquidus temperatures for olivine and plagioclase of a given basaltic composition: the phase with the highest liquidus temperature was the true liquidus phase as illustrated in the middle panel in Figure 1. Langmuir and Hanson (1981) and Nielsen (1988) modified this approach considerably by adding additional phases and temperature-composition algorithms for calculating the composition of the solid phases. Ghiorso et al. (1983) developed a regular solution model for basaltic liquids by extracting binary interaction coefficients from a wide range of experimentally produced pairs of crystal + glass (quenched liquid). This model lends itself to the prediction of liquidus equilibria by free energy minimization, as illustrated in the lower panel in Figure 1.

Each approach has its advantages and drawbacks. The major advantage of the graphical approach is its ability to inform the trained observer not only of the immediate liquidus phases of a composition but also of the proportions of important mineral components and of the course of crystallization. To overcome the shortcoming of a limited range of compositions to which these approaches apply, I began a program of parameterizing projected liquidus boundaries more than a decade ago and work is still in progress (Longhi, 1977, 1982, 1987a; Longhi and Pan, 1989). Then, as now, the conceptual framework is the set of quaternary liquidus equilibria in the system CaO-MgO-Al_2O_3-SiO_2 (CMAS) (Presnall et al., 1978; Longhi, 1987b), and the parameterizations account for the shifts in the liquidus boundaries produced by additional natural components such as FeO, TiO_2, and alkalies. Over the years these empirical expressions have become more complex

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Origin of Picritic Green Glass Magmas by Polybaric Fractional Fusion

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Calculations show that it is possible to generate picritic mare green glass compositions by inefficient polybaric fractional fusion of a low-Al_2O_3 source region. In these calculations it is assumed that a rising source region undergoes small amounts of pressure-release melting (1-1.2%/kbar) and that some of this melt escapes from the source into a reservoir that remains in close proximity to the source. At the end of the calculations, the pooled melt in the reservoir is a composite of small degrees of melt drawn off over a range of pressures from a progressively depleting source. The apparent pressure of melt segregation of the green glasses (≥20 kbar), derived from high-pressure melting experiments, is lower than the initial pressure of melting (~40 kbar) and higher than the final pressure of pooling (4-5 kbar). Polybaric fractional fusion thus overcomes some of the troublesome aspects of simpler static models, such as the need for high degrees of partial melting at lower pressures or modest degrees of partial melting at higher pressures followed by transit of the melt to the surface without fractionation. Varying degrees of efficiency of fractional fusion allow for a wide range of source compositions; however, the calculations show that in order for adiabatic decompression to supply sufficient heat to balance the heat of fusion, the melt extraction efficiency can be no more than ~0.4. In order to produce pooled melts in the Al_2O_3 concentration range of the green glasses (6-9 wt%) with inefficient fractional fusion plus small degrees of melting of the source, it is necessary to start with a very-low-Al_2O_3 source (≤1.5 wt%). Such a source is consistent with cumulate models of the mare basalt source region, but is lower in Al_2O_3 content than some estimates of the average lunar mantle (e.g., Taylor, 1982). Small variation of the ilmenite content of the source can produce pooled melts with a wide range of TiO_2 concentration; however, the high-Ti picritic glasses require a different or perhaps additional melting mechanism because they are too dense to have formed at pressures greater than 20-25 kbar and then risen to the surface (Delano, 1990). Two physical models of the melt reservoir seem possible. One is an initially barren olivine layer above the source into which the melt percolates. Calculations of the gravitational stability of melt in a porous layer suggest that the thickness of layer must be on the order of hundreds of meters to tens of kilometers in order for convective mixing (necessary for the success of the polybaric model) to take place. The other is a simple melt layer or bubble atop the rising source. Calculations of ascent velocities indicate that for a substantial range of density contrasts and partial melting the greater size of the source more than compensates for the lower density of the melt. Thus, melt may percolate out of the source, but free bodies of melt will not rise faster than the source, until the volume of the melt becomes a significant fraction of the source. The integrated degrees of melting required by the polybaric model (~12%) fall well within the no-escape range for reasonable choices of density contrasts. Although the high pressures of initial melting imply differentiated material at great depths (>1000 km) within the Moon, a shallower magma ocean is possible if radioactive heating of the layer beneath the magma ocean eventually leads to convective overturn.

INTRODUCTION

Picritic mare glasses are believed to have formed in widespread fire fountain eruptions (*Heiken et al.*, 1974). The compositions of these glasses span an impressive range in TiO_2 concentration (0.4 to more than 16 wt%; *Delano*, 1986) that exceeds the range observed in mare basalts. Several unanswered questions remain concerning the composition of their eruptive gas propellant (*Naughton et al.*, 1972; *Sato*, 1979), their relation to mare basalts (*Grove and Vaniman*, 1978; *Longhi*, 1987), and the depth of their source regions (*Delano*, 1980). Of these, the depth of the source regions is least amenable to a definitive answer. The purpose of this paper is to examine some of the existing models for the depths of the source regions and to propose a new one that is based in part on recent theoretical developments in magma transport. To simplify matters, I will concentrate on the glasses lowest in TiO_2, the so-called green glasses.

Traditionally, lunar petrologists have interpreted the point of olivine + pyroxene multiple saturation on the high-pressure liquidus of mare compositions, illustrated in Fig. 1, as the pressure of melt segregation (e.g., *Green et al.*, 1971; *Walker*

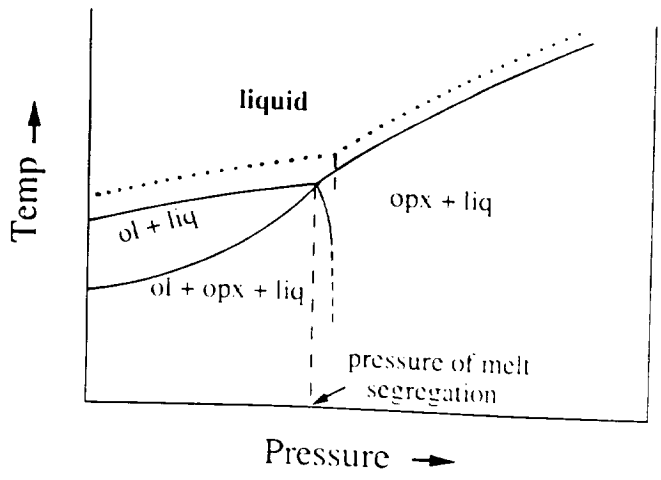

Fig. 1. Schematic pressure-temperature relationships for picritic green glasses. Dotted lines show the effect of increased olivine component. "Multiple saturation" refers to saturation of the melt with two or more phases. Abbreviations: ol = olivine; opx = orthopyroxene.

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Phase equilibria of dikes associated with Proterozoic anorthosite complexes

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ABSTRACT

We have investigated the petrogenesis of Proterozoic massif anorthosites through experimental phase equilibria studies of two compositions representative of intrusive bodies associated with anorthosite plutons, an anorthositic dike from the Nain Complex and an average high-Al gabbro composition from the Harp Lake Complex. Experiments on both compositions show that liquidus plagioclase becomes distinctly more albitic with increasing pressure. The anorthositic dike composition does not represent a liquid or a simple suspension of plagioclase in liquid because the phase assemblages and mineral compositions produced in the experiments do not match those in thin section. The discrepancy appears to be caused by large-scale heterogeneity in the dike sample, which is also evident in the whole set of anorthositic dike compositions and may be caused by open-system crystallization. The high-Al gabbro composition has plagioclase, orthopyroxene, and high-Ca pyroxene in its liquidus at 11.5 kbar. Orthopyroxene crystals formed at 10–11.5 kbar are similar in major and minor element composition to the most aluminous orthopyroxene megacrysts, and liquidus plagioclase compositions at 10–11.5 kbar overlap the bulk of plagioclase compositions reported for the Harp Lake Complex. These features are consistent with both the orthopyroxene megacrysts and much of the plagioclase crystallizing in lower-crustal or upper-mantle magma chambers from a parental magma similar to the average high-Al gabbro and then intruding upward as mushes or crystal-rich suspensions.

INTRODUCTION

The composition of magmas parental to Proterozoic massif anorthosites has eluded petrologists for a long time. As far back as 1917, N. L. Bowen noted that the fundamental problems of anorthosite genesis are finding a parent magma composition and determining a process by which the high modal proportions of plagioclase in the plutons could be produced. A number of models have been advanced over the years to explain the formation of anorthosites and associated granitic rocks from parental magmas ranging in composition from quartz dioritic (Green, 1969a, 1969b) to anorthositic or hyperfeldspathic (Buddington, 1939; Yoder and Tilley, 1962; Simmons and Hanson, 1978; Wiebe, 1979, 1980; Morse, 1982) to basaltic (Emslie, 1980; Longhi and Ashwal, 1985). There is little agreement on the crystallization conditions as well. Models range from high-pressure fractionation followed by diapiric rise of crystals or liquids (Emslie, 1980; Duchesne et al., 1985; Longhi and Ashwal, 1985) to in situ crystallization (Wiebe, 1979) to fractional crystallization (Green, 1969b). In this study we assess the feasibility of two potential parental magmas, anorthositic and basaltic liquids, by examining the phase equilibria of dikes associated with anorthosite plutons. We combined the results of this experimental study with other observations to assess a model for the genesis of massif anorthosites.

Geology

The Proterozoic massifs are the most voluminous of terrestrial anorthosites, with surface exposures ranging up to 30000 km^2 per complex. In detail, the complexes consist of numerous smaller plutons of variable size and composition (e.g., Emslie, 1980; Morse, 1982). Most massif anorthosites were emplaced between 1.2 and 1.7 b.y. and are associated with anorogenic magmatism or what has been interpreted as failed rifting (Morse, 1982; Emslie, 1985). Syenite, mangerite, and charnockite plutons and Fe-Ti oxide–rich rocks are spatially and temporally associated with many anorthosites. Field and geochemical evidence indicates that the silicic rocks are not comagmatic with the anorthosite plutons, although they are contemporaneous (e.g., Buddington, 1939; Ashwal and Seifert, 1980; Duchesne et al., 1985; Duchesne, 1990). This removes the constraint that anorthosite parent magmas must also be able to produce syenitic or granitic residual liquid through direct fractional crystallization (e.g., Bowen, 1917; Green, 1969b). It is also noteworthy that mafic and ultramafic rocks are not found in great abundance close to any anorthosites.

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Experimental petrology and petrogenesis of mare volcanics*

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Abstract—Mare volcanics consist of basalts and picritic pyroclastic glasses spanning a wide range of TiO_2 concentration. The more primitive low-Ti basalts and picritic glasses have olivine alone on their low-pressure liquidi. Most of the chemical variation among the low-Ti basalts is the result of olivine fractionation in a series of parental MgO-rich liquids differing in TiO_2 concentration. With one possible exception (Apollo 17 VLT) none of the picritic compositions is a suitable parent for any of the observed low-Ti basalts. Most of the chemical variation among the high-Ti basalts is the result of a series of magmas fractionated along the low-pressure olivine + armalcolite/ilmenite cotectic. All of the picritic high-Ti glasses have olivine alone on the liquidus, but none is a suitable parent for any of the basalts. Volcanics with intermediate TiO_2 concentrations (5 to 10 wt%) are widespread in the maria, even though they are not well represented in the sample collections; however, there is no evidence either among the samples or from remote sensing studies of basalts with >13 wt% TiO_2 that would be expected as differentiates of the picritic glasses with the highest TiO_2 concentrations.

Controlled-cooling-rate crystallization studies on a variety of mare compositions have provided the basis for reconstructing the size and, in some cases, stratigraphy of mare flows. Groundmass textures, crystal size, crystal morphology, nucleation density, and zoning patterns have all been employed to quantify cooling histories of mare basalts. A single-stage linear cooling rate may produce a porphyritic texture. Rapid cooling may also cause plagioclase to crystallize after a mineral that it precedes during slow cooling.

Mare basalts are highly reduced. Mineral assemblages and intrinsic oxygen fugacity measurements indicate f_{O_2} below the wüstite-iron buffer and at or near iron metal saturation. Accordingly, experiments run in high-purity iron capsules gained or lost little iron. Most basalts are undersaturated with respect to sulphur, so reduction through sulphur volatilization cannot be invoked to explain the presence of iron in olivine phenocrysts. The low oxidation state is most likely the result of melting a reduced interior under fluid-absent conditions. The progressive reduction of $Cr^{3+} \rightarrow Cr^{2+}$ and $Ti^{4+} \rightarrow Ti^{3+}$ in lunar melts permits the elimination of the Fe^{3+} that is present at iron-saturation in simple systems.

Crystal-liquid partition coefficients determined from melting experiments have been used in a wide range of calculations of major and trace element evolution. The coefficient for Fe-Mg exchange between olivine and liquid apparently varies with TiO_2 concentration of the liquid and is particularly useful in assessing whether fine-grained rocks have excess olivine. Nb and Ti have excess concentrations in mare basalts relative to adjacent REE in incompatibility diagrams. These excesses or positive anomalies are consistent with ilmenite accumulation in light of measured partition coefficients and imply continuous variation of accumulated ilmenite even in the low-Ti mare source regions.

Pressures of multiple saturation (olivine + pyroxene ± Cr-rich spinel ± ilmenite) are in the range of 5 to 12.5 kbar for primitive mare basalts and in the range of 18 to 25 kbar for the picritic glasses. Low-Ca pyroxene is the only pyroxene along the liquidus of the low-Ti basalts and glasses; however, augite is the pyroxene most commonly observed along the high-pressure liquidi of the high-Ti basalts: High-Ti picrites have augite in the subliquidus region at intermediate pressures where olivine is the liquidus phase, but orthopyroxene is the liquidus phase at multiple saturation. Because of the steep depth/pressure gradient in the outer portion of the Moon (20 km/kbar), these pressures imply: (a) great depths of melting within the Moon and (b) some means of transporting magmas hundreds of kilometers to the surface without significant chemical modification, if both olivine and pyroxene were left in the residuum. Modeling of major elements during polybaric partial melting suggests that it is possible to reproduce the composition and high-pressure signature of the low-Ti (green) picritic glasses by accumulating small degrees of melt extracted from an upwelling source region. In such a case, melting of the differentiated source must begin at >1000 km depth (40 kbar) and cease at ~100 km (5 kbar). If this model is correct, then experimental determination of the pressure of multiple saturation gives an average pressure of melting: the onset of melting is at higher pressure and actual segregation of the melt from the mantle is at lower pressure.

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INTRODUCTION

MARE VOLCANISM produced a diverse suite of basalts and volcanic glasses. Perhaps the most dramatic feature of these volcanics is the wide range of TiO_2 concentration (0.3 to

6. THE BULK COMPOSITION, MINERALOGY AND INTERNAL STRUCTURE OF MARS

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The only firm constraints on the bulk composition and internal structure of Mars are its size and mass. All previous estimates of Mars' composition and structure are based upon a cosmochemical model and an assumed value of the moment of inertia. However, the moment of inertia remains model dependent. The SNC meteorites (supposedly originating from Mars) are basaltic rocks that provide an independent means of estimating Mars' bulk composition and internal structure. Correlation of various pairs of elements is a means of estimating the ratios of key volatile, siderophile and chalcophile elements in the crust and mantle. By assuming approximately chondritic abundances for refractory elements (Al, Ca, Ti, Mg, Si, REE), Dreibus and Wänke (1985) have thus calculated absolute concentrations for a number of key elements. Their calculations show that the silicate portion of Mars is enriched in FeO, MnO, alkalis and halogens, but depleted in siderophile (Ni, Co) and chalcophile (Cu, Zn) elements relative to the Earth. These gross chemical characteristics of Mars are consistent with accretion from more volatile-rich material than the Earth, more extensive oxidation of Fe-metal, and separation of a Fe-Ni-S core. Mass balance considerations suggest a core that is 21.7% of the mass of Mars with 14.5 wt% S. Recalculation of the silicate portion into likely mineral components yields an uncompressed mantle density of 3.52 Mg m^{-3}. The uppermost Martian mantle is likely to be dominated by olivine and orthopyroxene, as is the Earth's upper mantle, although the Martian mantle has a lower MgO/(MgO + FeO) ratio (0.74 vs 0.89). The olivine-peridotite layer extends to a depth of 900 to 1100 km where the transition to silicate spinel begins. This transition is at a greater depth than in the Earth because of the shallower pressure gradient in Mars. The extent of the perovskite-wüstite-stishovite layer is highly dependent on the composition and size of the core. In the present model, there is a 200-km thick perovskite layer above the core-mantle boundary at 1700 km. The state of the core is unknown, but the absence of a strong magnetic field is consistent with weak convection, as would be expected in a totally molten or totally solid core that was not releasing latent heat. Calculations of the high-pressure liquidus and solidus temperatures indicate that for the case of a molten core the minimum temperature at the core-mantle boundary is ~ 2000 K, whereas for the case of a solid core the maximum temperature is ~ 1800 K. Summation of the masses in the various layers of Mars yields a value of 0.353 for the dimensionless moment of inertia, which is intermediate between the generally accepted value of 0.365 and the value of 0.345 predicated on a nonaxisymmetric distribution of mass about the Tharsis plateau.

I. INTRODUCTION

Our estimates of the nature and proportions of minerals in the Martian interior are based on three independent sets of observations: the Viking Lander soil analyses, the mean density and gravity field, and the compositions of the SNC (shergottites-nahklites-Chassigny) meteorites that we assume to be of Martian origin. These observations are the major parameters of models that produce a self-consistent picture of mass distribution, bulk composition and mineralogy.

Each set of observations is controversial in its own way: Do the soil compositions represent an average basalt composition or are they the result of extreme nonisochemical weathering of some now obscure protolith? How tightly do the mean density and gravity field constrain the mass distribution as expressed by the dimensionless moment of inertia? Do the SNC meteorites really come from Mars? The first question is discussed in chapters 18 and 19. The major arguments in favor of a Martian origin for the SNC's include a mineralogy distinct from lunar and other meteoritic basalts, relatively young crystallization ages (≤ 1.3 Gyr; summarized by McSween 1985), and a rare-gas isotopic signature in impact melt similar to that observed in the Martian atmosphere (Bogard and Johnson 1983; Becker and Pepin 1984). Furthermore, there are important similarities between the Viking soil analysis at Sandy Flats and the magmas from which the SNC's crystallized. Although we cannot be certain that the concentration of FeO in the Sandy Flats soil (> 20 wt% normalized; Toulmin et al. 1977) accurately reflects its basaltic

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Pressure effects, kinetics, and rheology of anorthositic and related magmas

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ABSTRACT

Anhydrous experiments on natural and synthetic starting materials with basaltic to anorthositic bulk compositions show a systematic increase in the albite component of near-liquidus plagioclase and in the Al_2O_3 content of orthopyroxene with increasing pressure. These results are consistent with crystallization of the highly aluminous orthopyroxene megacrysts and most of the plagioclase in massif anorthosite complexes at lower crustal pressures. Comparison of plagioclase compositions from near-liquidus and subliquidus experiments conducted in this laboratory with plagioclase compositions predicted at 1 atm for the experimental temperature and liquid compositions by various empirical models indicates that the shift to more albitic plagioclase is predominantly a pressure effect on the partitioning of albite (Ab) and anorthite (An) components between plagioclase and liquid. However, even when pressure terms are added to the models for Ab and An partitioning, there remain statistically significant compositional dependencies that are most apparent when the liquid composition is nepheline normative. These compositional dependencies probably arise from the absence of highly aluminous and nepheline-normative liquids in the data from which the models were constructed. Accordingly, we present empirical adjustments to the plagioclase-liquid models of Drake (1976), Weaver and Langmuir (1990), and Ariskin and Barmina (1990).

The positive pressure dependence of Al_2O_3 in orthopyroxene coexisting with plagioclase and liquid is almost entirely the result of changes in orthopyroxene-liquid partitioning and not related to increases in the Al_2O_3 concentration of the liquid. Data for Al_2O_3 partitioning from 46 orthopyroxene-liquid and 45 pigeonite-liquid pairs taken from the literature show that pressure is the most important control on the simple molar partition coefficient for Al_2O_3. Rapid crystal growth is rejected as an alternative explanation for the high Al_2O_3 contents of orthopyroxene megacrysts because rapid growth leads to low Cr_2O_3 concentrations in orthopyroxene, contrary to what is observed.

These results support polybaric models for massif anorthosite petrogenesis that entail accumulation of plagioclase in evolved basaltic magma chambers ponded in the lower crust followed by buoyant ascent of plagioclase-rich magmatic suspensions that intrude the upper crust, carrying rafts of orthopyroxene megacrysts. In thick, decompressing suspensions, the interplay of tie-line rotation and mass balance prevents plagioclase from becoming significantly more anorthitic. Experimental studies suggest that the transition from liquid- to solid-state rheology of plagioclase suspensions occurs at ~60% crystallinity for a homogeneous grain-size distribution and near static conditions. However, both motion of the suspension and uneven grain-size distribution shift the transition to higher crystallinities. Thus the transit of suspensions with leuconoritic composition (65–70% plagioclase) may be possible with minimal deformation of the entrained plagioclase. Formation of deformed anorthosite masses may then occur as second-stage buoyant segregations within the upper crustal magma chambers.

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SCIENTIFIC COMMENT

Terminology for trace-element partitioning

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Abstract—A self-consistent terminology for partitioning data is presented. Ratios of the concentration of a component in two phases are termed partition coefficients and given the symbol D. Ratios of partition coefficients are termed exchange coefficients and given the symbol K_D. The prefix "bulk" implies that these coefficients are weighted according to the proportions of coexisting phases. Bulk partition and bulk exchange coefficients are denoted by $\bar{D}$ and $\bar{K_D}$, respectively.

INTRODUCTION

THERE IS AN INCREASING interest in the data of element partitioning, both among geochemists and from those in other fields. This interest is reflected by the large number of partitioning studies published recently (see JONES, 1993, for a review). It has become clear that information from trace elements is able to place important constraints on magmatic processes and the chemical evolution of the Earth and other planets, and that a knowledge of partition coefficients is essential if these data are to be interpreted correctly.

Attempts are being made to integrate this information with that obtained using a wide range of other geochemical and geophysical approaches. It is therefore essential that partitioning information should be as accessible as possible to those from other fields. Unfortunately, the terminologies used in published partitioning studies are inconsistent and confused. The terms "partition coefficient" and "distribution coefficient" together with the symbols D, K_D, and K_d are generally used to denote various partitioning relations. However, the definitions of these terms and symbols varies greatly from one paper to another, causing much unnecessary confusion.

It is clear from the recent literature that there is an increasing need for a common terminology for these studies. This paper seeks to present such a terminology. We hope that its use will make the data of trace-element partitioning more accessible to a wide range of geoscientists. The terminology we present here is an extension of that suggested by TAKAHASHI and IRVINE (1981) and is consistent with that used by YARDLEY (1989) to describe partitioning relations in metamorphic systems. The use of this agreed terminology should help make the information obtained from trace-element geochemistry more accessible to the earth science community.

We hope that this terminology will become generally accepted and end the confusion that currently exists.

TERMINOLOGY AND DEFINITIONS

Although partition coefficients calculated on a molar basis provide the closest reflection of the thermodynamics that control partitioning, most trace-element models use partition coefficients calculated on a weight fraction basis. We therefore recommend that partition coefficients should be calculated from the weight fraction of the components involved unless otherwise stated. We suggest that the weight fraction of a component in a phase should be given the symbol C.

Simple concentration ratios between two phases are termed *partition coefficients*, in accordance with DENBIGH (1966), and will be given the symbol D; this is the "Teilungskoeffizient" of NERNST (1891). Subscripts will refer to the element of interest, and superscripts will refer to the phases concerned, with L being used for the melt phase. The phases between which the component partitions may be placed before the D where it is not convenient to superpose subscripts and superscripts. Thus, for component MO (where M is any cation) partitioning between phases α and β

$$C^{\alpha}_{MO}/C^{\beta}_{MO} = D^{\alpha/\beta}_{M}$$

$$= {}^{\alpha/\beta}D_{M}.$$

If partition coefficients are calculated on a molar basis, we recommend the term *molar partition coefficient* and the symbol D_*, with

Experimental study of a jotunite : constraints on the parent magma composition and crystallization conditions (P, T, fO_2) of the Bjerkreim-Sokndal layered intrusion (Norway).

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Abstract

The Bjerkreim-Sokndal layered intrusion is part of the Rogaland anorthosite Province of southern Norway and is made of cumulates of the anorthosite-mangerite-charnockite suite. This study presents experimental phase equilibrium data for one of the fine-grained jotunite (Tjörn locality) occurring along its northwestern lobe. These experimental data show that a jotunitic liquid similar in composition to the Tjörn jotunite, but slightly more magnesian and with a higher plagioclase component is the likely parent of macrocyclic units (MCU) III and IV of the intrusion. The limit of the olivine stability field in the experimentally determined phase diagram as well as comparison of the Al_2O_3 content of low-Ca pyroxenes from experiments and cumulates (≈ 1.5%) yield a pressure of emplacement ≤ 5 kb. Experimentally determined Fe -Ti oxide equilibria compared to the order of cumulus arrival in the intrusion show that the oxygen fugacity was close to FMQ during the early crystallization. It subsequently decreased relative to this buffer when magnetite disappeared from the cumulus assemblage and then increased until the reentry of this mineral. Calculated densities of experimental liquids show a density increase with fractionation at 7, 10 and 13 kb due to the predominance of plagioclase in the crystallizing assemblage. At 5kb and 1 atm (FMQ-1), where plagioclase is the liquidus phase, density first increases and then drops when olivine (5 kb) or olivine+ilmenite (1 atm : FMQ-1) precipitate. At 1 atm and NNO, the presence of both magnetite and ilmenite as near liquidus phases induces a density decrease. In the Bjerkreim magma chamber, oxides are early cumulus phases and liquid density is then supposed to have decreased during fractionation. This density path implies that new influxes of magma emplaced in the chamber were both hotter and denser than the resident magma. The density contrast inferred between plagioclase and the parent magma shows that this mineral was not able to sink in the magma, suggesting an *in situ* crystallization process.